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GLACIAL GEOLOGY
and the
Pleistocene Epoch

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Pleistocene Epoch

by Richard Foster Flint

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To the memory of

MAX DEMOREST

1910–1942

Outstanding glaciologist, excellent field companion, generous and thoughtful friend, who died to save the lives of others.

November 30, 1942

PREFACE

The Pleistocene epoch occupies a peculiarly important place in the time scale of geology, for it embraces the events of the latest million or more years in the history of the Earth and is therefore so recent that it bridges the gap between the geologic changes now in progress and the more remote past. "When the work of the geologist is finished," wrote Gilbert,[1] "and his final comprehensive report written, the longest and most important chapter will be upon the latest and shortest of the geologic periods. The chapter will be longest because the exceptional fullness of the record of the latest period will enable him to set forth most completely its complex history. The changes of each period — its erosion, its sedimentation, and its metamorphism — obliterate part of the records of its predecessor and of all earlier periods, so that the order of our knowledge must continue to be, as it now is, the inverse order of their antiquity."

This fact in itself furnishes an adequate reason for making the principal facts of the Pleistocene epoch compactly available, not only to geologists but also to ecologists, archeologists, geographers, and others whose studies reach back into the prehistoric realm. In addition, the increased pace of research upon Pleistocene problems in general, and problems in glacial geology in particular, that has been evident during the last two decades has emphasized the necessity, in this field, of a summary that will be at once a reference to the data already established and a means of indicating the areas and problems in which further research is most needed. These are the principal objectives of the present volume. No one knows better than its author that it falls short of attaining them. Knowledge of the Pleistocene has grown to such an extent that a complete reference work would become an encyclopedia. The consequent necessity for condensation has required the exercise of selective judgment at every turn. The list of references at the end of the book is far from complete, though an earnest effort has been made to see that it is representative. In particular it may lack important titles that have appeared in some countries during the war years and that have not yet been widely distributed.

[1] Gilbert 1890, p. 1.

This discussion treats the Pleistocene frankly from the point of view of glaciation, the outstanding characteristic that distinguishes the Pleistocene from the epochs that preceded it. The somewhat cumbersome title was selected with this fact in mind, in an effort not to create the impression that the work is a fully balanced treatment of every phase of the Pleistocene.

As is pointed out in Chapter 16, the correlations of Pleistocene events cited and suggested are, as far as possible, those based on geologic evidence rather than on archeologic evidence. In the presentation of geologic evidence itself stream-terrace data are used as little as possible in the belief that this class of data is more frequently subject to faulty interpretation than the data obtained from features of other kinds.

In particular this book avoids, in correlation, deduction from any theory of Pleistocene climatic fluctuation which sets up a fixed chronology of events. This conservative attitude is adopted on the principle that only when the stratigraphic column is built up strictly on geologic evidence can the influence of prejudice in favor of a particular theory of climate be avoided. To enable the reader to evaluate the reliability of the data used, a continuous effort has been made to discriminate between reasoning by induction from field evidence and reasoning by deduction from assumed general conditions.

The various aspects of the Pleistocene embrace many branches of geology and in addition other fields such as plant and animal ecology and phases of physics and astronomy. In consequence some of the data contained in this book are unavoidably scattered. It is impossible to assemble in a single chapter all the data pertinent to that chapter without unjustifiable repetition. The reader is therefore urged to make full use of the index, which has been compiled with this difficulty in mind.

References to literature are shown in footnotes, which lead to entries in the alphabetically arranged bibliography at the end of the book.

It would be impossible to mention the names of all those persons who have aided directly or indirectly in the preparation of this book, but my gratitude to them is none the less sincere. In particular I am indebted to colleagues and friends who have closely read various chapters and from whose expert criticism I have greatly benefited. Among these are Professors C. O. Dunbar, Adolph Knopf, and C. R. Longwell of Yale University, Dr. E. B. Knopf, Dr. A. L. Washburn, and Mr. Herbert G. Dorsey, Jr., of the U. S. Weather Bureau. In addition Professor G. E Hutchinson of the Department of Zoology, Professor Dirk Brouwer of the Department of Astronomy, and Professor Irving Rouse of the Department of Anthropology, all in Yale University, have helpfully discussed special problems.

In the heavy task of preparing and checking the bibliography I have had the expert help of Miss Gertrude Goodman, who also prepared the typescript for printing. The long and tedious business of reading typescript and proofs has been greatly lightened through the help of the quick and practiced eye of Margaret C. H. Flint. Finally I am indebted to several members of my current graduate class for reading proof on the typescript.

The courtesy of many persons and organizations who have furnished material for the illustrations is acknowledged beneath the illustrations themselves.

RICHARD FOSTER FLINT

NEW HAVEN, CONNECTICUT
December, 1945

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Chapter 1

THE ROOTS OF THE BASIC CONCEPTS

THE LAST MILLION YEARS

The last million years of geologic history have witnessed changes in the physical aspect of the Earth and in the distribution of animals and plants on the Earth's surface such as are not recorded in any earlier span of time of comparable length. Similar events have occurred in the more remote geologic past, but the record of them is dim and will never furnish more than a small fraction of the details which, though very incomplete, have already been drawn from a study of the Pleistocene epoch.

A rich store of information has been accumulated, most of it within little more than a century, on this latest of the geologic epochs. A far richer store still awaits the inquiring eye of the geologist, the zoologist, the botanist, and the anthropologist. All this knowledge, actual and potential, is made possible by the recency of the sedimentary deposits, the fossil animals and plants, the multitudinous land forms, and the geologic structures that constitute the record. These features are so young that the slow geologic processes which in the course of time hide such things from view — erosion, burial, submergence, distortion, metamorphism — have destroyed them only in part or not at all. In 1839 Lyell named this epoch Pleistocene — *most recent* — because of the modern aspect of the invertebrate fossils in its deposits, but if he had had full knowledge then of all the other lines of evidence he could not have named the unit more appropriately.

In most minds the Pleistocene connotes glaciation, and properly so, because glaciation was the outstanding event of this million-year epoch. This is true even though very extensive glaciers covered the lands during less than half of this span of time. Hence glaciation, its causes, its direct and indirect effects, and the criteria of its recognition form the principal part of the account that follows. But glaciation was only one result of the climatic changes that set off the Pleistocene from the epoch that preceded it. Accordingly the other results are successively discussed, but since many of them are direct results of glaciation and only indirect results of climatic change they are treated afterward. Directly or

indirectly the Pleistocene climates constitute the background of the entire discussion.

THE GLACIAL THEORY[1]

The idea that glaciers were formerly far more extensive than at present dates only from the second quarter of the nineteenth century, though it antedates the recognition and naming of the Pleistocene epoch. Long before that time Swiss peasants, who lived their lives in valleys and on mountain sides freshly abandoned by Alpine glaciers, had drawn the fundamental inference from the evidence that lay everywhere about them. The scored and polished bosses of bedrock, projecting through the turf of Alpine meadows, were unmistakable. The profusion of smoothed and faceted boulders spread wide through the fields and in the forests, and the ridgelike abandoned end moraines on the valley floors, were identical with those at the margins of the glaciers themselves and could hardly be misinterpreted by one who could see both boulders and glaciers in a single view.

The inference had been drawn by Kuhn, Saussure, and Hugi before the nineteenth century began. It was first expressed in detail by a Swiss civil engineer named Venetz, who in 1821 read a paper before the Society of Natural History (the Helvetic Society) at Luzern in which he argued that the glaciers of the Alps had at some former time been expanded on an enormous scale.[2] Three years later, in 1824, Jens Esmark, in Norway, reached a similar conclusion concerning the glaciers in the mountains of Norway.[3]

Venetz, like the generations of peasants before him, seems not at first to have projected his imagination beyond the Alps. However, in 1829, after longer reflection and probably a good deal of discussion with his scientific friends, he stated the belief that not only the Alps but the plains north of them, and the whole of northern Europe as well, had once been glaciated.

A colleague in the Helvetic Society, Jean de Charpentier, though at first incredulous, was stimulated to further observation, and in 1834 he read a paper strongly supporting Venetz's views and strengthening their proof.

John Louis Rodolphe Agassiz, a young zoologist and a member of the Helvetic Society, frankly doubted so sweeping an inference, and in the summer of 1836 he arranged to visit the Diablerets glaciers and the

[1] See the historical summaries in North 1943; McCallien 1941; Merrill 1906.

[2] Venetz acknowledged an unpublished inference of this nature made by J. P. Perraudin in 1815.

[3] Esmark 1824.

moraines of the Rhône Valley with Charpentier, primarily in order to convince his friend of his error. But after some weeks of study and discussion in the neighborhood of Bex, it was Agassiz who was convinced. He realized not only that Venetz and Charpentier were right, but also that they had not gone far enough. At some time between the dawning of this conviction and the next year, 1837, when Agassiz addressed the Helvetic Society, the concept of an ice age emerged. For the address pictured "a great ice period" caused by climatic changes and marked by a vast sheet of ice extending from the North Pole to the Alps and to central Asia—all *before* the Alpine region had been lifted up to mountainous heights.[4] Not until ten years later did he recognize that the former glaciers of northern Europe were quite separate from the former Alpine glaciers, and that the Alpine glaciers postdated, rather than antedated, the making of the Alps themselves.[5]

One of the facts that led Agassiz to extend the ice-age concept to Asia as well as Europe was the occurrence, well known even at that time, of extinct mammoths and other animals in frozen soil in northern Siberia.

In his general concept Agassiz was anticipated by a professor in the Academy of Forestry at Dreissigacker named Bernhardi, who had written in the *Jahrbuch für Mineralogie* that the distribution of moraines and *Findlinge*—erratic boulders—compelled the belief that glacier ice from the north polar region had once extended as far south as Germany.[6] Bernhardi knew of the views of both Esmark and Venetz, but it required a lively imagination to bridge the gap between Germany and the Arctic with glacier ice. However, no one in Germany paid serious attention to this idea (despite Agassiz's conversions in Britain and America) until 1875, when Otto Torell, a Swedish geologist, brought to Berlin overwhelming evidence that Bernhardi had been right, and that to him rightfully belonged the distinction of having been the first to recognize an ice age.

In a sense Bernhardi deserves more credit for his contribution to the concept of an ice age than do any of the three Swiss. They had living glaciers before them and could make direct comparison between areas now occupied by ice and those evidently abandoned by ice. Bernhardi's evidence was less fresh, less distinct, less abundant, and, above all, geographically removed from direct comparison with the work of living glaciers.

[4] The substance of the address was published three years later (Agassiz 1840), a year before the amplified views of Charpentier (1841) were issued in book form. Charpentier felt that Agassiz should have deferred publication until after the appearance of his own book.

[5] Agassiz 1847.

[6] Bernhardi 1832.

The recognition of far-traveled erratic stones was not new. Some years before 1787 that keen observer, Saussure, had recognized boulders resting on the slopes of the Jura Mountains yet clearly derived from the Alps fifty miles to the south. Saussure noted them, but he thought that they had got there as a result of a cataclysmic flood.[7] It was James Hutton of Edinburgh who, reading Saussure a few years later, first reasoned that those boulders must have been glacier-borne to their positions.[8] Hutton anticipated even Venetz and Bernhardi.

The geologists of the British Isles, like Bernhardi, worked under the disadvantage of being far from any existing glaciers. They suffered under the additional handicap of an insular position (a handicap evident in other Britons both before and since that time). Never far from the sea and its implications, these men (Hutton excepted) interpreted their erratic boulders and other glacial deposits as the product of a vast though short-lived submergence beneath the sea — some thought of it as a sudden rush of water and others as the biblical Deluge — during which icebergs dropped the foreign rocks upon the surface. The concept of floating ice was promoted by an interest in polar exploration that was then rapidly expanding. It is because of this belief in submergence that glacial deposits collectively were called drift, a term that has survived its abandoned context by a hundred years. Even Sir Charles Lyell, that master of inductive reasoning, was misled by the evidence.

The greatest proponent of the submergence concept had been William Buckland, professor of geology at Oxford.[9] Interested by the new discoveries in Switzerland, he visited Agassiz in 1838 and soon realized that the British and Alpine evidences were similar. Convinced that his own earlier interpretation would no longer stand and that the Swiss were on the right track, Buckland persuaded Agassiz to visit the United Kingdom in 1840, and there accompanied him into the field. Although warmly supported by Buckland and in part by Lyell,[10] the conclusion reached by Agassiz that Britain had been glaciated was at first ridiculed by many British geologists. The concept of a marine origin of the drift lingered in many minds until it was firmly dispelled by two classic papers, one published in 1862 by T. F. Jamieson;[11] the other in 1863 by Archibald Geikie.[12] These papers mark a turning point in the development of the study of glacial geology. During the decades of controversy

[7] Saussure 1786–1796, p. 202.

[8] Playfair 1802, sec. 349.

[9] Buckland 1823.

[10] Lyell recognized that the moraines of the Scottish Highlands proved former glaciers, but for England generally he held to the iceberg theory.

[11] Jamieson 1862.

[12] A. Geikie 1863.

real study had stagnated. New investigation took a new lease on life and a new epoch of discovery began. Venetz, Charpentier, Agassiz, and Bernhardi were established as the founders of the theory of the glacial epoch, though Hutton and Esmark, at least, anticipated them as being among the founders of glacial research.

Whatever unpublished thoughts about the origin of the drift may have been entertained in America before 1846, the numerous published ideas, except one, were strongly influenced by religious bias or by British ideas of submergence.[13] The exception was a carefully thought-out inductive statement, based strictly on field observation, by Peter Dobson,[14] a Connecticut mill owner. Dobson reached an erroneous conclusion, that the facets and scratches on the stones in glacial deposits had been made as the stones were dragged over the ground by floating icebergs; but he reached it by the right method.

No sooner had the concept of the glacial origin of the drift, published as the result of Buckland's and Agassiz's 1840 trip through Scotland, reached America than it found a single ready response. Edward Hitchcock, in a first annual address[15] before the newly formed Association of American Geologists, strongly supported Agassiz, though later, curiously enough, he partly recanted.

Eighteen forty-six was the year in which Agassiz himself arrived in America to become a professor at Harvard. His presence in America undoubtedly hastened the wide acceptance of the glacial theory, but with many, belief in the iceberg theory lasted as long as it did in Britain. Slowly the glacial view progressed against resistance; one of the best American statements in advocacy of it was made by J. D. Dana.[16] The last scientific opposition to it in North America died in 1899 with J. W. Dawson; in England Sir Henry Howorth published a 1500-page argument in opposition as late as 1905.

In 1854, before the glacial theory had become widely accepted, evidence was uncovered both in Britain and in the Alps that showed that the spread of glaciers over great areas of low lands was not a single event but had been repeated. Gradually indications accumulated that within the limits of the Pleistocene epoch there had been four distinct glacial ages, when glaciers spread over nearly a third of the world's land area, separated by interglacial times of deglaciation — warmer times when the extent of glaciers was greatly restricted. Whether other, lesser glacial ages preceded the earliest of the four recognized cold times or

[13] See Merrill 1906.
[14] Dobson 1826.
[15] E. Hitchcock 1841.
[16] Dana 1863, pp. 541–546.

intervened between them is a matter for future discovery. For the present it is enough to say that the concept of repeated glaciation grew up with the glacial theory itself.

THE PLUVIAL LAKES

Hardly had the fact of widespread glaciation become generally established when the effect of the glacial climates on the saline lakes in arid regions was perceived. In 1863 T. F. Jamieson, writing in Edinburgh, said: "Now this heat and dryness [of the arid regions] being much lessened during the glacial period, there must have resulted a much smaller evaporation, which would no longer balance the inflow. These lakes would therefore swell and rise in level."[17] He mentioned specifically some of the great saline lakes of Asia — the Caspian, Aral, Balkhash, and Lop-Nor lakes.

Meanwhile Lartet published the results of studies he had just completed in the region of the Dead Sea, another great salt lake. He had found deposits, evidently made by the lake, high up above its present shores, and he inferred that the lake had expanded during an ice age, which was clearly recorded by glacial features on Mt. Lebanon, west of the Dead Sea.[18] Lartet's inference, which Jamieson's deduction anticipated by two years, was later confirmed and established. Today it is widely evident that the glacial ages were marked in many dry regions by the expansion of lakes, which shrank during the interglacial times. Except for repeated studies of the Aral-Caspian system, study of this problem in the Old World has lagged. The most detailed and reliable evidence now available comes from western North America, yet even in that region a vast deal still remains to be done.

Although Jamieson deduced that arid-basin lakes should have expanded during a glacial age, and although Lartet established that both glaciation and lake expansion had in fact occurred in the Dead Sea region, proof of the simultaneous occurrence of both events was not furnished until much later. In his classic study of ancient Lake Bonneville in Utah Gilbert showed that end moraines built by valley glaciers descending from the Wasatch Mountains lie below the highest shoreline made by the lake, though the exact relation of moraines to strandline was obscure.[19] Still earlier, Russell had demonstrated, through the relation of the shoreline of the expanded former Mono Lake in eastern California to moraines of a Sierra Nevada glacier, that lake and glacier were essentially contemporaneous.[20]

[17] Jamieson 1863, p. 258.
[18] Lartet 1865, p. 798.
[19] Gilbert 1890, p. 318.
[20] I. C. Russell 1889, p. 369; see also Blackwelder 1931, p. 889.

It is unlikely that a similar relationship can be demonstrated for most lakes, but the evidence of their former expansion — in some cases two or three expansions — is so well-nigh universal that few students of the Pleistocene have serious doubts that lakes in the dry regions of all continents expanded and shrank synchronously with the onset and waning of the successive glacial ages, or that both lakes and glaciers were the result of a single underlying cause — climatic fluctuation. Because of the increased rainfall that must have characterized the now-dry regions when the lakes were expanded, the climatic conditions that obtained there and then are often referred to as *pluvial,* and the extensive high-water times have been called *pluvial ages.*

THE SWINGING SEALEVEL

The glacial (pluvial) ages witnessed not only vast enlargements of glaciers in high and middle latitudes and widespread growth of lakes in dry continental regions but also worldwide emergence of shallow continental shelves as the level of the sea was conspicuously lowered.

In 1842, while the glacial theory was still being hotly debated, Charles Maclaren deduced that, if the glacialists were right in supposing a former vast expansion of ice on the lands, the level of the sea must then have been hundreds of feet lower than now.[21] Much later it was recognized that the sealevels of the interglacial ages must, in some cases at least, have been higher than those of today. The amount of water substance at the Earth's surface is nearly constant; when glaciers and lakes are formed on the lands, the seawater must diminish by a corresponding amount. When the glaciers melt and the lakes dry up, the water is returned to the sea.

On steep, rugged coasts the resulting changes in land areas were small. But on the broad continental shelves comparatively small reductions in sealevel resulted in the addition of large areas to the adjacent lands and in corresponding local changes in climate and in forced migrations of animals and plants. The fluctuation of sealevel, synchronously with the coming and going of the glacial ages, is a relationship now universally acknowledged.

THE FAILING CRUST BENEATH THE ICE SHEETS

As early as the sixteenth century (and probably long before then) inhabitants of the northern Baltic region had noticed that their coasts were rising, in relation to sealevel, at a rate that made the change clearly

[21] Maclaren 1842, p. 365.

apparent within a very few generations. Much later the presence in this region, high above the sea, of sediments containing marine invertebrates, seals, and whales was recognized. This fact gradually came to be connected with the coastal changes as evidence of upwarping of the Earth's crust.

The true cause of the upwarping was first perceived in 1865 by Jamieson, who had been impressed by the recognition of similar marine sediments lying above sealevel along the Scottish coast. So weak and flexible was the crust, he believed, that the weight of accumulated glacier ice had caused it to subside; when the ice melted away the crust slowly rose and regained its former position.[22] This view, according to which the northern Baltic region has not yet recovered its preglacial altitude and is still rising, has been universally adopted. Among its early proponents were Whittlesey[23] and Shaler.[24] A test of the principle involved was made by Gilbert,[25] who attributed the domelike warping of strandlines of glacial Lake Bonneville to removal of the very considerable weight of its water when the lake dried up.

Detailed studies of the glaciated regions of both Europe and North America have shown that the form of the unwarping, both past and present, is domelike, and that the centers of the warped areas lie in the central regions of the former ice sheets.

As the great glaciers thickened and spread, the crust slowly subsided under their weight, and as the ice melted, the deformed crust gradually rose to assume its former shape. This subsidence and elevation on a majestic scale must have taken place with the waxing and waning of each glacial age, synchronously with the fall and rise of the surface of the sea.

THE LATE-CENOZOIC MOUNTAINS

No one who examines the present-day distribution of glaciers can fail to realize that glaciers are related to highlands. Without high and extensive mountains, some of them situated in the paths of moist winds, extensive glaciation can not occur.

It has long been known that at present there are more extensive mountains and much higher general continental altitudes than have existed throughout the greater part of clearly recorded geologic time. But only recently has it become apparent that the uplifts and mountain-making movements that created the present high lands are largely of

[22] Jamieson 1865.
[23] Whittlesey 1868.
[24] Shaler 1874, p. 335.
[25] Gilbert 1890, pp. 362–383, pl. 50.

post-Miocene date, and that mountain uplifts amounting to many thousands of feet have occurred within the Pleistocene epoch itself. The Cordilleran mountain system in both North and South America, the Alps-Caucasus-Central Asian system, and many others have attained the greater part of their present height during the late Cenozoic. But more significantly for general glaciation the mountains of Scandinavia, Greenland, and Labrador, which are believed to have given origin to some of the largest of the ice sheets, are now believed to have reached their present heights in very late Cenozoic time.

This widespread series of uplifts, taken as a whole, set the stage for the glaciers that were the natural response to the Pleistocene temperature reductions at the Earth's surface. Without the highlands the chilling of the Earth could at best have created only glaciers of limited extent in very high latitudes.

The building of the late-Cenozoic highlands, then, was a long and essential prologue to the building of the late-Cenozoic glaciers.

PLEISTOCENE STRATIGRAPHY BEYOND THE GLACIATED REGIONS

While the glacial theory was taking form in Switzerland and Germany, geologists were studying the Cenozoic strata in the Paris basin and were subdividing them on a basis of the invertebrate fossils they contained. One of the results of this work was the recognition and naming of the *Pleistocene series*.[26] Thus two separate lines of inquiry, one into glacial geology and one into paleontology, were being pushed forward simultaneously by two independent groups, both of whom were at first oblivious of the very close interrelation these two channels of research would prove to have.

Little by little the relation of glacial to nonglacial deposits around the borders of the glaciated regions has begun to emerge, though we are still far from an adequate correlation of glacial strata with contemporaneous marine and terrestrial sediments laid down remote from the territories covered by the ice sheets. The objective view of the Pleistocene epoch must consider the Earth's surface, giving due weight to the events remote from glaciation and perhaps little related to it, as well as to the more unusual effects of the glaciers themselves. It is very probable that because of the worldwide results of the Pleistocene climatic changes the relation of glacial to nonglacial features will prove to be much closer than is even now believed.

[26] Lyell 1839, p. 621.

THE MIGRATIONS OF ANIMALS AND PLANTS

From the widely scattered information on the animals and plants collected from Pleistocene strata there has begun to emerge a picture, dim as yet but significant. Comparatively few new forms of life appeared during the Pleistocene epoch, which was characterized rather by repeated migrations of entire floras and faunas. The evidence consists of two or more assemblages of fossils in different strata in a single locality, each assemblage representing markedly different climatic conditions. Relationships such as these seem to indicate clearly the mass movements of animals and plants, as changing climates drove them toward the equator and again permitted them gradually to move into higher latitudes. This concept fits into the emerging picture of the Pleistocene as a time of fluctuating temperature, with conditions in the middle latitudes now colder and now warmer than they are today.

THE OUTLINES OF THE UNIFIED PICTURE

When these various groups of facts and the basic concepts derived from them are considered together, they form a consistent though incomplete picture. If world events during the later Cenozoic could somehow be shown in generalized form on motion-picture film, and the action enormously speeded up, the synchronous operation of the great movements involved could be quickly perceived. The scene would begin with the gradual unheaval of the lands and the raising of high mountain chains. As the temperature became lower and again higher the most conspicuous effect would be the waxing and waning of glaciers. Accompanying this would be the subsidence and later recovery of the crust beneath them, the slow fall and rise of the surface of the sea, and, in opposite phase, the growth and shrinkage of the pluvial lakes on all the continents. Moving in harmony with these changes in inanimate things would be the irregular processions of animals and plants driven from their habitats by temperature changes, by inundations, by desiccations, or by the incursions of the ice sheets.

This telluric symphony, in four movements (and perhaps others that we have not yet perceived) is still being played. Even the earlier movements are so recent that much evidence of them is still legible. This evidence has made it clear to us that the events of the Pleistocene epoch, in their great intensity and wide extent, have been unlike anything appearing in the record of Earth history throughout the span of the 200 million preceding years. It has shown that, at least once during the Pleistocene epoch, glaciers have covered nearly one-third of the land area of the world. It has demonstrated that the million years, more or less, of Pleistocene time have embraced several glacial ages. These cold

times were separated by interglacial ages, longer than the glacial ages, during which the glaciers dwindled or disappeared in climates as warm as the present climates, or even warmer.

The facts of the Pleistocene combine to pose a climatic problem of the first magnitude. If we could solve the problem of the glacial climates and the longer-enduring interglacial climates we would have taken a long step forward in understanding the climatic changes recorded in sedimentary rocks far older than the Pleistocene. Also we would understand much better the meaning of the less spectacular yet distinct climatic changes that have occurred since the beginning of human history.

One of the main objectives of the study of the Pleistocene epoch is historical. It is the reconstruction, step by step, of the climatic events and their sequels that have affected the world during the last million years. The reconstruction follows the classic method of geologists — inference from the evidence of former events, and comparison with events now taking place.

The principal part of our discussion is concerned with glaciation. In the main the evidence from which we draw our inferences is, naturally, evidence of former glaciers. But it includes also features, made at the same times as the glaciations, by streams and lakes, by the wind, and by the sea. Many of these features are so closely related to the glaciations that the glaciations can not be fully understood without reference to them.

In addition to inference from physical evidence we have a second line of attack. We can study existing glaciers and from them draw analogies, comparing their behavior with what we have inferred concerning the behavior of greater glaciers now vanished. Armed with knowledge of the glaciers of today we can deduce the events of earlier times when glaciers, though they did not differ physically from those we now see, were far more extensive than now. Analogy alone is not a reliable means of reconstructing an entire glacial age, but, when used in conjunction with inference drawn from the geologic record, analogy has great value.

We shall begin, therefore, with the basic facts about the glaciers that exist today and then turn to the geologic record of glaciers that have vanished. In other words we shall begin with facts from *glaciology*[27]

[27] These terms are sometimes misused. A *glaciologist* is a scientist whose special field of study is the physical structure and deformation of glacier ice. A *glacial geologist* is a geologist whose special study is the geology of glacial features. *Glacialist* is the term used in Britain, during the three or four decades following the announcement of the glacial theory, for an individual who supported the theory that the glacial drift was the product of former glaciers, as opposed to those who believed it to be a deposit made in the sea or during the biblical Deluge. In more recent times this term has been used incorrectly to mean *glacial geologist,* though fortunately such improper usage is now almost extinct.

— the study of the physics of glaciers — and shall continue with facts from *glacial geology* — the geology of glacial features — which is a very different thing. This sequence is adopted in the belief that, if certain facts of glaciology are understood, the significance of a host of geologic features made by glaciers becomes much more clear.

Still later we shall discuss the crustal movements, the swinging sea-level, the pluvial lakes and other features of the nonglaciated regions, and the migrations of living things necessitated by the slow shifting of the climatic zones. Last but not least we shall attempt to evaluate what is known and what has been speculated about the fundamental causes of the climatic oscillations that mark the Pleistocene epoch as unique among the divisions of later geologic time.

Chapter 2

GLACIER ICE AND GLACIER MOTION

BASIC TYPES OF GLACIERS

Fundamentally and in the simplest view there are two kinds of glaciers, valley glaciers and ice sheets. *Valley glaciers* are streams of ice that flow downward through valleys in highlands. Like streams of water they may be short or long, wide or narrow, single or with branching tributaries. The lengths of some are measured in hundreds of yards; the lengths of others, in scores of miles. The longest valley glacier yet measured is the west branch of Hubbard Glacier in Alaska, 75 miles overall. Common to lofty highlands in many parts of the world, they have also been called *mountain glaciers* and *ice streams*. Very small ones have been called *cirque glaciers* and *glacierets*. Some originate in small snowfields hardly wider than the glaciers themselves. Others, *outlet glaciers,* serve as escape drainageways for ice sheets dammed up behind mountain barriers, as around the margins of Greenland and the Antarctic Continent. But all belong to the general valley-glacier type. As they flow down the (usually steep) slopes, they mold themselves to the shape of the underlying ground and thus come to follow pre-existing valleys.

Ice sheets, in contrast, are not confined to valleys but are broad, cake-like ice masses usually (though not invariably) occupying highlands. Small ice sheets, and sometimes large ones as well, are referred to as *ice caps*. Very small thin ice caps lie on the broad uplands of the Norwegian mountains. Larger ones are found in Iceland, on islands in the Arctic Sea such as Spitsbergen and the Franz Josef Land archipelago, and on North American islands such as Baffin and Ellesmere islands.

Far larger are the ice sheet of Greenland, 637,000 square miles in area, and the vast Antarctic Ice Sheet, with an area of about 5,000,000 square miles. These two ice masses are thick enough to bury entire mountain ranges. Together with their smaller counterparts they flow outward literally *under their own weight*. The pressure caused by the great weight of the overlying ice[1] squeezes the basal ice slowly outward.

Intermediate between valley glaciers and ice sheets are the *piedmont*

[1] Equivalent to about 30 tons per square foot per 1000 feet of thickness.

glaciers that occupy broad lowlands at the bases of steep mountain slopes.[2] Each is the spread-out, expanded terminal part of a valley glacier descending the highland, or the coalesced combination of two, three, or many parallel valley glaciers. The large Malaspina and Bering glaciers on the coast of Alaska at latitude 60° and the Frederikshaab Glacier on the west coast of Greenland are examples.

The differences among valley glaciers, piedmont glaciers, and ice sheets result mainly from the fact that ice flows and is therefore able to adjust or mold itself to the irregularities of the ground on which it lies. However, the ice in all glaciers is the same, and a brief discussion of it is necessary before we examine in more detail the nature of existing glaciers.

PHYSICAL CHARACTERISTICS OF GLACIER ICE

Glaciology, the science of existing glaciers, is fundamentally a study of the deformation of glacier ice. To understand the nature of the deformation of any solid substance, whether ice or any other solid, involves complicated inquiry, laboratory experimentation, and mathematical analysis and is difficult in the extreme. Research into the deformation of ice is only in its infancy, and, although some general principles are known, important questions remain to be answered.

Our present purpose is to draw upon what is generally known about the physics of glacier ice only to the small extent necessary to create a background for an understanding of the work of glaciers in former times. Therefore we consciously avoid discussing the many fascinating structural features of glaciers and confine ourselves to general principles.[3]

Ice is both a mineral and a rock, which under the temperatures at the Earth's surface is exceptionally perishable. Ice comes into existence wherever water freezes — in streams, lakes and the sea, in the atmosphere, and in the soil; and the resulting kinds of ice have distinctive characteristics. The most distinctive and in many ways the most remarkable kind is *glacier ice.* Glacier ice is not simply a rock. It is distinctly a metamorphic rock. It can be described as *firm ice made by the recrystallization of a mass of fallen snow.*

The progressive evolution of snowflakes through more or less loose granular ice (*névé* or *firn*) into firm glacier ice is a matter of common observation and has long been recognized.[4] The first stage of this evolution, the change from a bank of freshly fallen, flaky snow into a

[2] Piedmont glaciers that occupy a trough or basin fed by valley glaciers from mountains on two or more sides are termed *intermont glaciers* by Matthes (1942, p. 152).

[3] For further material on strictly glaciologic features see the extensive, up-to-date, and thoughtfully selected list of 104 titles prepared by Matthes (1942, pp. 215–219).

[4] See an excellent treatment in Ahlmann 1935.

mass of loose, granular névé, easily takes place within a single season. Two processes are involved: sublimation and a melting-refreezing process. By sublimation, molecules of water vapor escape, chiefly from the edges of snowflakes, and attach themselves to the central parts of other snowflakes in such a way as to maintain the hexagonal crystalline structure originally possessed by the flakes. Sublimation thus converts flakes into crystalline granules, each usually a fraction of a millimeter in diameter.

Before the sublimation process has come to an end, the growing granules begin to melt at the points of their mutual contact, because pressure there is slightly increased; and the meltwater descends and refreezes, still preserving the crystalline form of each granule. Granules that are initially large grow at the expense of adjacent smaller ones. The result is that the grains in any mass of névé tend to become about equal in size as they gradually increase in average diameter. Thus the granules of névé, a small fraction of a millimeter in diameter when first converted from snowflakes, grow to a millimeter or so during the course of a single season. This process of growth and change is known as *recrystallization*.

The second and final stage in the transition is the conversion of névé into glacier ice. This process is slow, requiring not one season but a long time. During this time the growing mass receives an ever-thickening cover of snow which in turn is continuously converted into névé. The weight of this cover constitutes a load, which causes settling and compaction of the névé underneath and gradual expulsion of the air between the grains of névé. Thus little by little the ice changes from a loose pile of granules having a porosity of nearly 50 per cent to a continuous solid. All that remain of the extensive pore space that characterized the névé are bubbles of trapped air. The continuous solid is glacier ice, which ordinarily does not come into existence until its substance has become buried to a depth of at least 100 feet. Viewed under the microscope the constituent grains of this new glacier ice are a few millimeters in diameter and subequal in size. Further changes beyond this stage are caused by deformation of the ice. This takes place mainly by flow, a natural result of the pressure resulting from the piling up of additional snow on top of the mass.

GENERAL NATURE OF GLACIER MOTION[5]

At the moment when flow begins, the mass of ice becomes a glacier. This we can describe as *a body of ice (usually with some névé) consisting of recrystallized snow, lying wholly or largely on land, and showing*

[5] A good technical treatise is Lagally 1934.

evidence of present or former flow. The rather cumbersome qualifications are necessary because the ice of a glacier hardly ever is wholly separable from the névé into which it grades, because the terminal parts of some glaciers are afloat, and because some glaciers, although not now flowing, show by their metamorphic structure that they have flowed extensively in the recent past (Fig. 5).

In some respects a large body of glacier ice behaves like a body of water. But one difference in particular is noteworthy. When released from the confining pressure of a container, the body of water will immediately change its shape by flowing. On the other hand ice, in common with all solids, possesses some rigidity. In consequence a body of ice, when released from confining pressure, will begin to flow only after the application of force sufficient to overcome its rigidity. If the ice body is thick enough, the pressure caused by its own weight will start the process of flow. Once the ice has begun to flow it apparently obeys the laws of fluid mechanics. Therefore we say that glacier ice "flows" even though it is not, properly speaking, a fluid substance.

It has been rather substantially confirmed that the flow of glacier ice takes place chiefly through continuous changes in the crystalline grains of which the ice is composed. Under the stresses to which ice in the glacier is constantly subject, individual crystals grow in size, are deformed (apparently by a pack-of-cards-like gliding of thin plates parallel with the basal planes of the crystals), and rotate with respect to neighboring crystals. The result of these changes is streamline flow in the glacier, "a process of differential shearing in which infinitesimal 'layers' slide past one another on planes that are theoretically infinite in number. Traces of these planes are the lines of flow or streamlines that mark paths along which infinitesimal particles of the flowing substance are moved."[6] The lines of flow would be straight and parallel were it not for differences of velocity caused by physical obstructions and other factors. In consequence the lines of flow are commonly curves.

Probably the rate of flow of a mass of glacier ice increases as the square of the depth. In the higher and outer parts of a glacier, the ice is under such slight confining pressure that it is brittle and is deformed by fracture rather than by flow. The most obvious examples of this superficial fracture are the crevasses that cut the surfaces of some parts of nearly every glacier. In places in this outer zone rigid masses of ice are displaced along well-defined surfaces, some of which are true thrust faults, whereas others are marked by thin layers of clear ice — the "blue bands" of dozens of reports and studies — in which extensive recrystallization has taken place.

[6] Demorest 1942, pp. 31–32.

A glacier, like the Earth's crust itself, can be said to possess an outer rigid shell — the zone of fracture — and an inner mobile mass — the zone of flow.[7] Under certain stress conditions, rupture localized along definite planes can and does occur at depth, but as a rule fracture is confined to the superficial zone.

The crevasses and other less conspicuous fractures that characterize the predominantly rigid upper zone of a glacier are believed to extend down to depths of 100 to 200 feet, below which increased pressure causes the ice to flow together and close up all openings. Actual observations on which this depth figure is based are scanty, but they are consistent. Matthes[8] calculated from observations in the Bighorn Mountains that on a 12 per cent grade a mass of névé must be built up to a thickness of 125 feet before it can begin to flow. Demorest[9] calculated that a glacier occupying a broad gently sloping floor in the Glacier Park District in Montana had begun to flow when the thickness of névé had reached 150 feet. According to Bowman[10] névé piled up on a horizontal surface will begin to flow when its thickness has reached about 250 feet. From seismic tests carried out on the South Crillon and Klooch glaciers in Alaska, R. P. Goldthwait[11] inferred that the maximum depth of the crevasses in those glaciers is about 100 feet. This depth would vary somewhat with varying temperatures.

The crevasses to which the brittle surface zone of a glacier is subject are primarily the result of tension, and they occur most commonly at places where comparatively thin ice is flowing over an abruptly steepened slope. The broad central areas of ice sheets, underlain by thick ice and having a very gentle surface slope, are characteristically free of crevasses. Only near the ice-sheet margins, where thinner ice passes over rock bosses or escapes down steep valleys in mountainous terrain, are crevasses abundant.

The actual process of flow can not be seen in a glacier because the flowing part is always concealed beneath 100 to 200 feet of overlying ice. However, in the glacier's terminal zone, melting exposes to view ice that is now brittle but that was formerly deep in the body of the glacier. This ice shows by its foliated, metamorphic structure that it has been subjected to flow (Fig. 5). This ice is truly a metamorphic rock, like the more durable metamorphic rocks that have lain within the zone of flow in the Earth's crust and have become exposed to our view through deep,

[7] This seems to have been pointed out first by Tarr and Martin (1914, p. 187). See also Hess 1933, pp. 80–81, 92–93; von Engeln 1934, p. 401.

[8] Matthes 1900, p. 190.

[9] Demorest 1938, p. 724.

[10] Bowman 1916, p. 293.

[11] R. P. Goldthwait 1936.

long-continued erosion. Similar ice has been produced in the laboratory by subjecting undeformed ice to high pressures.

Because velocity increases with increasing depth, we can picture the general distribution of movement in an ideal ice sheet. Apparently in a thick circular ice sheet lying on a horizontal floor the mobile ice near the base of the glacier is squeezed out or extruded from beneath the less mobile ice near the top. Thus the most rapid flow occurs near the base of the ice (but not at the base, because there movement is slowed by frictional drag along the floor beneath the glacier). In the central area of the ice sheet the movement of the ice must be mainly downward. Gradually, however, it must change direction away from the center along any radius, until in the outer part of the glacier the movement is mainly outward. The concept of squeezing out of the lower part of the glacier is not wholly a deductive one. Observations made during a period of 20 years on a glacier in the Alps, whose névé-covered head occupies a flat or basinlike area, show that the motion in the head of this glacier must be distributed in this general manner.[12]

In a thin glacier flowing down a sloping floor, such as a valley floor, movement is affected by another gravitative factor resulting from the slope of the floor. At first thought the flow of such a glacier might seem to be analogous to the flow of a stream of water. But at least one conspicuous difference exists in the presence of the upper zone of fracture, which rides as a less mobile mass on the more mobile ice beneath it and is carried downstream by this ice underneath.

The long profiles of the floors of many valley glaciers consist of steep pitches alternating with flats and even basins. In such glaciers the streamlines of flow near the base of the ice must be inclined downward over the steep pitches and upward over the reverse slopes of the basins. In fact ice, when it is thick enough to flow at all, molds itself to the minutest topographic details of the surface over which it passes, flowing around and over obstacles only a few inches high. This is shown, in areas recently uncovered by wasting ice, by the relation of small scratches, made by stones carried in the base of the ice, to minor obstacles.[13]

Repeated measurements with stakes driven into the ice at intervals across valley glaciers have shown that the surface ice moves faster along the center line of the glacier than it does near the sides. In this respect the valley glacier is like a stream of water. These measurements apply, of course, to the upper zone of fracture, which is carried forward mainly by the drag of the more mobile ice beneath it and therefore reflects to some extent the rate of flow of the ice on which it rides. The greater

[12] Streiff-Becker 1938.
[13] Demorest 1938.

velocity along the central line of the valley is a result primarily of the greater thickness of the ice along that line and secondarily of frictional drag of the lateral ice along the valley sideslopes.

In any glacier the decreased thickness at its terminus, brought about by wastage as explained in Chapter 3, creates an obstruction to the flow of the ice behind it. This is because mobility diminishes with diminishing thickness, and therefore, other factors equal, the ice can not flow as rapidly near the terminus as it can at some distance upstream from the terminus. In fact, where the ice is so thin that the superficial zone of fracture extends from the glacier's upper surface all the way down to the subglacial floor, flow can not occur at all. The inert ice forms an obstruction to the flowing ice behind it. The ice from upstream crowds against it and moves forward and in the direction of easiest escape. If the rate of flow is small it may be counterbalanced by wastage at the surface and may die out upward. But strong flow can cause thrust faults in the brittle ice above and can bulge up the surface of the ice. Very strong flow can break through the zone of fracture and override the motionless or nearly motionless ice that forms the obstruction. The terminal zones of glaciers testify to this predominantly forward and upward motion by both blue bands and definite thrust planes, both of which are oriented in this direction.[14]

Apparently the obstruction formed by the terminal ice is the chief reason why the long profiles of glaciers steepen near their termini. The obstruction causes the flowing ice to build up the slope to a profile steep enough to permit the obstruction to be overcome. The greater the slope of the subglacial floor, the less steep the upper surface of the ice need be. In consequence the termini of glaciers occupying steeply sloping valleys show minimum steepening of their upper surfaces.

With this brief introduction we turn to the regimen, balance sheet, or economy of a glacier in an attempt to make clear what keeps a glacier going.

[14] Cf. Slater 1925.

Chapter 3

REGIMEN OF GLACIERS

INTRODUCTION

Save under very unusual circumstances a glacier is not a static body. It is in a condition of continual change, thickening or thinning, increasing or decreasing in area, sensitively responding to the slight changes in climate that continually affect it. No two successive years are exactly alike in the amount of snow that falls or in the amount of heat that reaches the Earth's surface at any place. In their sensitive reaction to climatic differences in successive years, decades, and centuries, glaciers examined over periods of many years are perhaps the most delicate climatic indicators to which we have access. An extreme example[1] is the Black Rapids Glacier, 45 miles south of Big Delta, Alaska. In 1937 this glacier was 10 miles long. It occupies a valley that drains a large upland basin filled with névé and ice. For many years before 1936 this glacier had been slowly shrinking. Probably late in 1936 it began to expand, and within a period of four or five months it had lengthened by 3 miles. During this time its terminus crept forward at the record average rate of 115 feet per day. Later investigation brought out the fact that during the period 1929–1932 the snowfall in this district had been exceptionally great. The accumulation of snow in the upland basin is thought to have accelerated the rate of flow of the Black Rapids outlet glacier, which drained the basin, and seven years after the beginning of the heavy snowfall this acceleration made itself felt at the terminus of the outlet 10 miles away.

The Black Rapids Glacier represents a very short-term change; yet such short-term changes must have affected thousands of glaciers thousands of times within the million-year span of the Pleistocene epoch. Other fluctuations of longer duration are in the record. Since the latter part of the nineteenth century glaciers all over the world, with local and temporary exceptions, have been shrinking. In some regions the shrinkage has been little short of catastrophic. This condition of wastage can only be a consequence of worldwide climatic change (decreased snowfall or increased average summer temperature or both), small in amount but nevertheless great enough to produce a response in the glaciers of every

[1] Hance 1937.

continent. If long continued, this climatic change probably would result in the disappearance of glaciers throughout the world. If reversed, the trend would cause general expansion of the glaciers and would produce conditions like those at the heights of the glacial ages when large parts of the land area of the world were buried beneath ice.

FACTORS IN THE REGIMEN

We are now in a position to analyze the factors that together control the regimen of any glacier. In a mass of snow or névé that is too thin to flow these factors are two: (1) *nourishment* by snowfall, resulting in the *accumulation* of more snow, and (2) melting and evaporation (together constituting the process of *ablation*), resulting in losses of snow or névé. If annual ablation (usually occurring mainly in summer) just balances annual accumulation (usually occurring mainly in winter) the net change in the amount of snow or névé on the ground is nil. An excess of annual ablation over annual accumulation during a long enough period will culminate in the shrinkage and disappearance of the mass. On the other hand a preponderance of snowfall over ablation will produce net accumulation, and consequent expansion of the mass of névé until it is thick enough to induce flow. At this moment the mass of névé becomes a glacier and a new factor is introduced into the regimen: the transfer of ice from one part of the glacier to another by flow. The greater the excess of nourishment over ablation, the more rapid the rate of flow. Furthermore, in the glacier we have a loss factor not operative in the névé mass—*calving*. Where the terminus of a glacier lies in water, the most important terminal loss is usually not through ablation but through the calving (breaking off and floating away) of large pieces of ice, often many hundreds of feet in diameter. The rapidity of calving as compared with ablation is demonstrated by the termini of broad glaciers that lie partly in standing water and partly on land. The terminal sectors that are afloat invariably show more rapid recession than those sectors which rest on land. Although characteristic of glaciers that end in water, calving occurs also in glaciers that end on land, provided their floors slope outward very steeply.[2] The three processes by which glaciers lose substance—melting, evaporation, and calving—are together spoken of as *wastage*.

CALVING

Some further information about calving should be recorded. Where a glacier ends in water, melting is more rapid at the waterline than either

[2] Cf. Bretz 1935, p. 197, and legend to fig. 285.

above or below it, owing to the higher mean temperature of the surface water than that of the deeper water or of the air. In consequence a notch or groove is melted into the ice at the waterline, leaving the higher ice overhanging while the ice well below the waterline juts as a sloping apron far out beyond the notch. The same notching is common in bergs and in sea ice.

The terminal part of every glacier is greatly fractured by the stresses incidental to flow. The presence of fractures, many of them vertical or near-vertical, coupled with the presence of the waterline notch, sets up the conditions under which calving takes place. Two kinds of calving are seen at glacier termini that are bathed in water. (1) Blocklike pieces of the overhanging front break off, of their own weight, along fractures and fall or slide into the water. (2) Pieces of the jutting underwater apron break free along fractures and are buoyed up to a floating position.

Although most bergs are small, many attain lengths of several hundred feet. One berg 7 miles by 3½ miles by 50 feet is on record, and bergs rising 450 feet above the waterline have been measured.[3] The visible part of a floating berg represents only a fraction of its actual volume. Owing to the fact that the specific gravity of glacier ice, although variable, is not much less than that of water, 85 to 90 per cent of the volume of the berg is submerged.

Because of the great number of nearly vertical fractures in the terminal parts of most glaciers, calving from above the waterline results in the maintenance of a steep cliff. This cliff, characteristic of glaciers ending in water, may reach heights of 500 feet.

Calving is more efficient than terminal ablation because it removes ice from the terminal zone of a glacier more rapidly. In consequence a broad ice front resting partly on low land and partly in the water of a bay shows a curving re-entrant where it crosses the bay. The re-entrant is greatest where the water is deepest (Fig. 7).

Where fractures are closely spaced, calving takes place about as rapidly as the glacier, moving into deepening water, is buoyed up by flotation. In consequence the termini of most fractured calving glaciers probably are aground. Bottom soundings can rarely be taken near a terminal ice cliff because underwater calving is unpredictable and is very unhealthy for boats, even of large size. However, if the glacier is aground, the depth of water must be less than 7 to 9 times the height of the terminal cliff.

In rare instances the terminal zone of a glacier may be afloat throughout a wide area. The most striking example is the Ross Shelf Ice in the Ross Sea, Antarctica, believed to be afloat throughout an area of perhaps

[3] E. H. Smith 1931, p. 107.

150,000 square miles. It is relatively free of fractures, and although thin (the terminal cliff rises 50 to 200 feet above sealevel) it is maintained partly by the accumulation of snow on its own surface.

Most bergs consist of almost pure ice and include little or no rock material. However, a small proportion of bergs contain accumulations, commonly in the form of crude layers, of rock fragments. As the berg melts, its rock content is dropped to the bottom, where it mingles with the other sediments of the sea or lake floor. Although the average amount of sediment per berg thus transported and deposited is small, the accumulation of berg-rafted sediments on wide lake- and sea-floor areas must be very large.

MUTUAL RELATIONS OF NOURISHMENT, WASTAGE, AND FLOW

In a Valley Glacier

The surface of an ideal valley glacier that drains out of a small mountain snowfield can be divided into two distinct parts: (1) An upstream *area of accumulation*. Over this area the annual snowfall is not entirely removed by ablation, so that substance is added to this part of the glacier year by year. (2) A downstream *area of ablation*. This area loses not only all the snow that falls directly upon it but also some of the underlying glacier ice as well. In consequence it undergoes a net loss of substance year by year.

The area of ablation is easily recognized at the height of the summer season when ablation is at its maximum, because at that time it has lost the covering of snow it acquired during the winter, and bare ice is exposed at its surface. In contrast the area of accumulation is covered with snow or névé throughout the entire year. The line on the surface of the glacier that separates these two areas is the *névé line* (also called *firn line*). It is determinable only at the height of summer, when ablation has reached its maximum and before the snow has again begun to fall. Its position fluctuates somewhat from one year to the next, but under stable conditions it tends to be about the same from year to year.

From beneath the area of accumulation ice is transferred by flow to the area of ablation, making good the loss caused by ablation in this area. If snowfall (controlled in most regions by winter conditions) and ablation (controlled in most regions by summer temperature and summer cloudiness) were to remain unchanged from year to year, the downstream transfer of ice by flow would be uniform, the position of the glacier terminus would be fixed, and the regimen of the glacier would be *in equilibrium*. However, equilibrium in a glacier is not often attained, and it is never maintained over a long period of years — prin-

cipally because snowfall varies from one year to the next. To be sure, ablation varies also, but there is no doubt that variations in nourishment are a far more fundamental cause of most fluctuations in glaciers than variations in ablation.

Thus a long succession of unusually snowy winters will thicken the ice in the area of accumulation. As the excess flows away, the rate of flow of the glacier will increase, just as the velocity of a river increases in time of flood. And if the rate of ablation downstream remains unchanged the glacier will expand, becoming both thicker and longer.

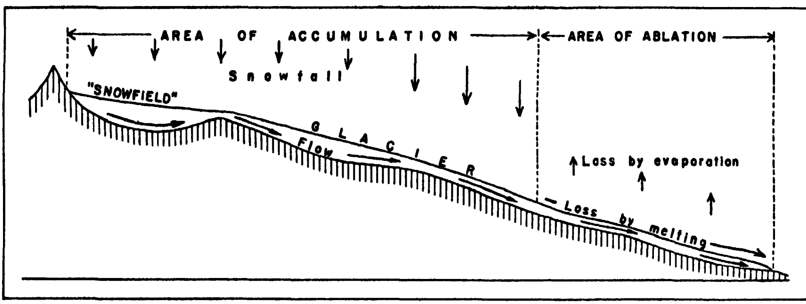


FIG. 1. Elements in the regimen of an ideal valley glacier.

Conversely a long succession of open winters will thin the ice in the area of accumulation, the rate of flow will diminish correspondingly, and, because normal losses in the area of ablation are not made good by increments of ice at the customary rate, the glacier shrinks, becoming both thinner and shorter. If thinning proceeds to an extreme degree, the ice is no longer able to flow and becomes motionless — stagnant. To put it in another way, the zone of flow in the ice thins until it disappears, thus bringing the base of the brittle zone of fracture down to the ground.

This regimen is much like that of a stream of water that drains a shallow lake, flows down a mountain valley into a desert, and there disappears by evaporation. The three factors of rainfall, stream flow, and evaporation are so closely interrelated that even a slight change in one factor affects both the others. In the valley glacier, accumulation, flow, and wastage are similarly interrelated (Fig. 1).

In an Ice Sheet

The regimen of an ice sheet is not as simple as that of a valley glacier, and it requires a separate explanation. It is easiest to visualize if we first think of a circular ice sheet on which accumulation is assumed to be

greatest at the center, whereas ablation is greatest at the periphery. The difference between net accumulation at the center and net wastage at the periphery is compensated by flow of ice from the central area downward and outward to the periphery. If the three factors of accumulation, wastage, and flow were exactly balanced, the regimen of the glacier

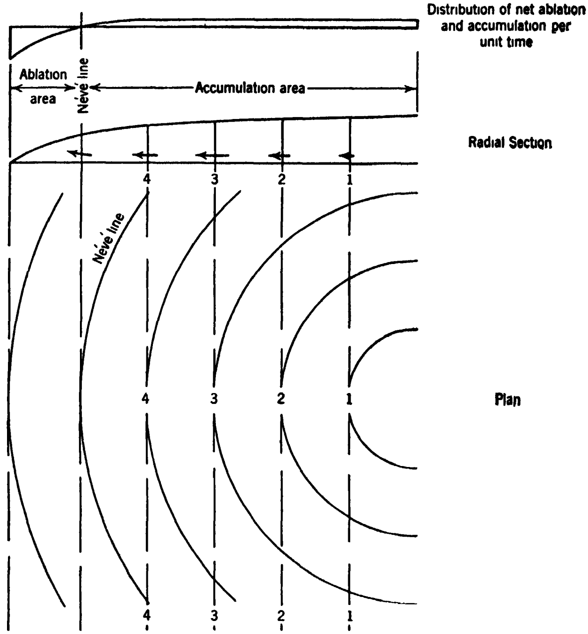


FIG. 2. Plan and radial section of ideal circular ice sheet (suggestive only; not quantitative). (Modified after Demorest.)

Upper figure shows general distribution of net accumulation (above the névé line) and net ablation (below the névé line) in a unit of time.

Radial section suggests, by means of arrows, relative rates of flow at various distances from center of ice sheet.

Plan shows, by arcs, successive circular vertical cross sections of the ice sheet.

would be in equilibrium. The ice in the central part would be kept at a constant thickness by continued accumulation, the periphery would be renewed by flow of ice at the exact rate at which it was wasted by ablation, and at any given point the rate of flow would be constant.

These statements appear to be true of an ideal circular ice sheet in equilibrium[4] (Fig. 2):

1. *The velocity of flow* through any concentric circular vertical section

[4] Demorest 1942, pp. 40–43.

is approximately proportional to the distance from the center of the glacier.

2. *The surface slope* (which determines the chief pressure differential in the ice beneath) is proportional to the velocity and is therefore also *proportional to the distance from the center of the ice sheet.*

3. It follows from statement 2 that *ice thickness must decrease outward from center to periphery.* Average mobility therefore decreases outward, and an additionally steepened surface slope is necessary in order to maintain the flow discharge of ice. Because of two distinct factors, therefore, this slope must steepen outward from the center at an increasing rate. In other words, *the surface must be convex up.*

Consider next a circular ice sheet on which maximum accumulation

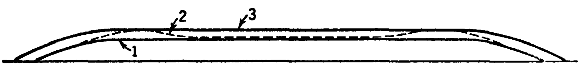


FIG. 3. Growth of an ideal circular ice sheet with maximum accumulation in a marginal belt. (Demorest.)

Profile 1. Condition of equilibrium.

Profile 2. Profile resulting from increased accumulation in marginal belt.

Profile 3. Equilibrium re-established by flow both outward and inward from beneath the belt of maximum accumulation.

is distributed evenly around a marginal belt rather than at the center. If this ice sheet is in equilibrium, its surface everywhere inside the belt of maximum accumulation will be horizontal (Fig. 3). Outside this belt there will be radial flow. Any increase in the rate of accumulation that is sufficient to thicken the ice sheet will be effective first in the circular belt of maximum accumulation. Here the surface will slope outward to the periphery and also inward toward the center. The inward slope will set up, in the ice beneath it, flow toward the center of the ice sheet. Convergence toward the center will produce an obstructing effect, with oblique-upward movement of ice. Rise of the surface of the ice sheet will result, and this will continue until the surface of the ice sheet once more becomes nearly horizontal throughout its entire extent, thus re-establishing equilibrium.

Regardless of how accumulation is distributed over an ice sheet, movements tending toward the establishment of equilibrium will take place according to the principles implied in the two ideal examples just cited. Probably no actual ice sheet has ever been nourished with maximum accumulation either at the center or around the entire margin. Probably some small ice sheets have received more or less uniform nourishment over their entire surfaces, and some large ones have almost certainly received maximum nourishment through considerable sectors of their

margins. The surfaces of these latter ice sheets should have sloped gently away from the sectors being actively nourished toward the sectors receiving little nourishment. The ice sheets would have been asymmetrical in vertical profile as they undoubtedly were also in ground plan.

Increase of accumulation without a corresponding increase of ablation would thicken the glacier, would increase the rate of flow of the ice, and would cause the ice sheet as a whole to spread outward and increase in area — to *expand*. Decrease of accumulation without change in ablation would produce the opposite effect: it would reduce both the thickness and the area of the ice sheet, which could be said to *shrink*. Thinning of the ice (likely to take place only in the peripheral area) sufficient to cause disappearance of the zone of flow would leave only the zone of fracture, which would lie motionless and stagnant.

RATES OF FLOW

Rates of flow vary with so many variable controlling factors that they are of little practical significance. The flow of the surface ice of valley glaciers has been measured hundreds of times in favorable localities by driving stakes into the surface of the glacier and then determining with surveying instruments the successive positions of the stakes at stated intervals of time. Frequently the results can be expressed in inches per day, though some rapidly flowing valley glaciers have shown measured rates of several tens of feet per day. The rate of flow of any glacier, measured over a long time as has been done in the Alps, varies from one period of years to the next; sometimes even from one year to the next. Nearly all glaciers show seasonal variation because the temperature affects the rate of flow of the ice at the surface.

It was pointed out earlier that the rate of flow of a glacier, as measured on the surface, is distinctly greater along the center line (where the flowing ice is thickest and least subject to frictional drag against the valley sides) than at the lateral margins. Usually also the rate of flow is greater throughout the long middle part of the glacier than near either the head or the terminus. As a rule the gentle slope and large cross-sectional area of the snowfield at the head prevent rapid flow there, while increasing ablation toward the terminus thins the glacier and reduces its mobility to such an extent that flow is reduced. At the extreme lower tip of the glacier there is no flow at all: the steeply convex upper surface of the ice here brings the zone of fracture down to the ground and pinches out the zone of flow entirely. Accordingly this small part of the glacier is either pushed forward bodily by the flowing ice behind it or is overriden by the flowing ice.

The flow of ice sheets can not be measured so easily because maximum flow occurs only at depth, beyond our reach. Demorest[5] estimated that even the maximum rates of flow near the margin, the most rapidly flowing part, of an ice sheet "are not greater than a few inches or fractions of an inch per day." Inward the rate diminishes to the center, where necessarily the only motion is mainly downward.

The highest rates yet measured come from the terminal parts of large outlet glaciers of the Greenland Ice Sheet. The Karajak outlet has yielded daily rates, in the summer season, approaching 50 feet, and the Storström outlet, during a year of measurement, yielded a rate of 5610 feet. These high rates of flow result from the great ice reservoir of the Greenland Ice Sheet whose enormous weight squeezes out the basal ice into the confined channels of the outlets with rapidity.

RELATION OF THINNING TO TERMINAL RETREAT

From the time when glacier regimens first began to be studied up to about 1930, the shrinkage of a glacier was thought of almost entirely in terms of retreat (recession, inward migration) of the glacier's terminus. Retreat of the terminus was obvious, easily seen, and easily measured, and the less obvious third dimension, thickness, was generally neglected.

However, painstaking year-to-year measurements have shown that in a shrinking glacier the area of ablation, below and beyond the névé line, undergoes thinning which adds up to conspicuous volume losses. For example, on the Nisqually Glacier on Mt. Rainier, Washington, accurate measurements made on a segment extending from the terminus through 3000 feet upstream show that during the 30-year period 1910–1940 the mean annual recession of the terminus was more than 70 feet, but the mean annual loss by thinning was 6.6 feet. Similarly, on the Hintereisferner, a glacier in the eastern Alps, during the 24-year period 1894–1917 the total terminal recession was 1006 feet, but the total thinning in the terminal zone amounted to 340 feet. In both these examples the amount of retreat exceeds the amount of thinning at any one place, but the thinning figure is an average applying to a large area, within the terminal zone, which is much greater than the area uncovered by the retreating terminus.[6] Thus in both examples far more ice was lost

[5] Demorest 1942, p. 54.

[6] Ablation is caused by conduction of heat to the ice surface, not only from warm air but also from warm rain. In regions of maritime climate, such as Iceland and southern Greenland, rain may, within a limited period, cause more ablation than does warm air.

by thinning of the entire glacier below the névé line than by inward migration of the terminus itself. A simple calculation from the results of the detailed measurements of ablation on the Fourteenth of July Glacier in northwestern Spitsbergen made by Ahlmann in the summer of 1934[7] shows that, of the total ablation measured, about 99 per cent of the volume of ice lost by ablation resulted from thinning of the terminal zone and only about 1 per cent was referable to retreat of the terminus.

In fact it seems probable that retreat of the terminus of a shrinking glacier is the result far less of wastage concentrated at the margin than of general thinning throughout the terminal zone. This is brought out in Fig. 4, which shows successive surfaces of a large outlet glacier

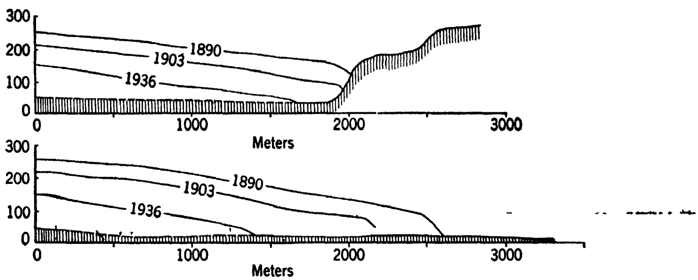


FIG. 4. Successive partial cross sections (above) and long sections (below) of the Hoffelsjokull, a large outlet glacier in southern Iceland, measured about 1890, and again in 1903 and 1936, showing progressive thinning. (After Ahlmann and Thorarinsson.)

measured at respective intervals of 13 and 33 years. These surfaces are *essentially parallel* through a distance of 6000 to 8000 feet back from the terminus.

Now ablation not only thins the ice over a wide terminal area; it also reduces the amount of "new" ice brought in by flow to counterbalance ablation losses, and it reduces the rate of flow, which, as we have seen, is controlled by ice thickness. If, as seems probable, rate of flow varies approximately as the square of the thickness, then a small decrease in thickness will result in a much larger decrease in rate of flow. As flow becomes slower, ablation increases still more. This spiral of cause and effect means, simply, that, when a glacier begins to shrink, thinning in the terminal zone (and, secondarily, retreat of the terminus) takes place at an accelerating rate. Regional thinning over a wide area is shown by

[7] Ahlmann 1935.

U. S. Navy Alaskan Aerial Survey Expedition

FIG. 5. Wood Glacier, Geikie Inlet, Glacier Bay district, Alaska, as it appeared in 1929, a separated residual mass of ice about 2 square miles in area.

comparative surveys of the many glaciers of the Glacier Bay district of Alaska.[8]

As thinning proceeds, the rate of flow progressively diminishes. In the Eliot Glacier on Mt. Hood, Oregon, the movement at a measured transverse line in 1925–1928 averaged 10.3 feet per year, but by 1943, after considerable thinning, it had slowed down to 4.7 feet.[9] In parts of the terminal zone the ice becomes so thin that the zone of flow described in Chapter 2 is pinched out, leaving the zone of fracture extending right down to the ground. Here the ice is stagnant (dead) and devoid of further flow. Where drastic thinning occurs over a wide terminal belt of ice that overlies irregular ground, the high ground gradually appears above the surface of the thinning ice, just as islands appear in a lake whose level is falling. The progressive emergence of these islands (or

[8] Flint 1942, p. 114.
[9] Matthes and Phillips 1943.

nunataks[10]) has been measured in Iceland glaciers through many decades. As the nunataks enlarge and coalesce, parts of the glacier that occupy valleys and basins become separated from the rest of the glacier (Fig. 5). Whether they are stagnant or not depends on their thickness. Other ways in which ice may become detached or may stagnate are discussed elsewhere.[11]

[10] This Eskimo word for mountains that project through the Greenland Ice Sheet has come into wide use as a name for all features of this kind, regardless of size or location.
[11] Flint 1942.

Chapter 4

FORM AND DISTRIBUTION OF EXISTING GLACIERS

REGIONAL SNOWLINE[1]

In high latitudes and at high altitudes there are many areas in which fallen snow is not entirely melted during the summer season but in part persists to go through the changes into névé and glacier ice already described. These areas are aptly termed *regions of snow accumulation,* for in them under present climatic conditions there is a net accumulation of snow. The line that separates them from the far more extensive areas in lower latitudes and at lower altitudes in which the snow disappears in summer is the *regional snowline.*[2] On an ice sheet, of course, the regional snowline coincides with the névé line and can be fixed quite accurately, provided an observer is on hand to make the measurement in late summer, the only time of year at which the determination can be made. On the other hand, as nearly all mountains and other highlands are marked by deep valleys and intervening ridges oriented in various directions, the regional snowline in them can never be determined exactly; it can be fixed only approximately, within hardly less than a few hundred feet.[3] Thus in most regions this "line" is really not a line at all. It is a zone (and a very irregular zone at that) in which snow sheltered from both wind and sun may be at much lower altitudes than snow exposed to the wind or to rapid ablation.[4] Yet, however irregular locally because of topographic and other influences, the altitude of the regional snowline changes consistently through long distances, forming a sort of world pattern. The height of the regional snowline is controlled mainly by temperature (especially summer temperature) and secondarily by the abundance of snowfall. In the broad view, therefore, the snowline is low near the poles where low temperatures, especially low summer temperatures, inhibit ablation, and high in equatorial latitudes where ablation is severe. At the equator, however, the snowline is not as high as in the dry horse latitudes — roughly 20° to 30°

[1] An excellent treatment of this subject is given by Matthes 1942.

[2] Sometimes also called *climatic snowline.*

[3] Various methods of approximating the regional snowline are set forth in Dainelli and Marinelli 1928, pp. 86–98.

[4] The lower limit of this zone is essentially what has been called the *orographic snowline.* As pointed out in Chapter 6, it is the feature that is recorded roughly by the floors of cirques and is somewhat lower than the climatic snowline.

both north and south of the equator — especially to the east of broad oceanic areas. Thus on Mts. Kenya, Ruwenzori, and Kilimanjaro in East Africa, all very close to the equator, the snowline ranges between 15,000 and 17,800 feet; in southern Tibet, at nearly 30° north latitude, it is more than 20,000 feet; and in the Andes between 20° and 25° south latitude it is more than 21,000 feet.

Regardless of latitude, abundant snowfall depresses the snowline. Thus on Mt. Olympus, on the peninsula between Puget Sound and the Pacific, with an annual precipitation of 150 inches, the snowline stands at 6000 feet, whereas on Mt. Rainier, in nearly the same latitude but farther inland and with an annual precipitation of 100 inches, the snowline stands at more than 11,000 feet.

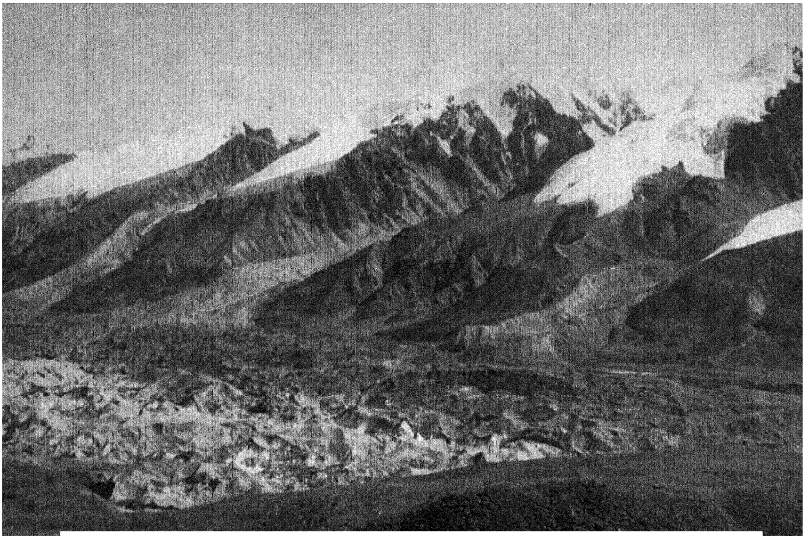
The general descent of the snowline from the horse latitudes toward the poles is very gradual at first. But in latitude 35° to 50° in both hemispheres it steepens conspicuously. The probable explanation is important, because it has a bearing on the former greater glaciation of both hemispheres.

There is an important relationship between the regional snowline and the vertical distribution of snowfall on a highland.[5] Where the highland has sufficient altitude, the mean annual snowfall increases upward to a maximum value and then, at higher altitudes, diminishes. Under present climatic conditions in low and middle latitudes, the snowline lies higher than the level of maximum snowfall, but in the polar regions the snowline lies below that level. In other words the snowline declines toward the poles at a steeper angle than the level of maximum snowfall. This has the result that as the snowline descends it successively occupies positions at which there is more snow and thus increases the persistence of fallen snow into and through the summer season. Where the snowline intersects the level of maximum snowfall there is a startling increase in the amount of snow remaining on the ground and, in especially favorable places, in the amount of glacier ice. The two lines cross in the high-middle latitudes, 50° to 65°. Here the present-day snowline bends more steeply downward toward the poles, and here, in North America, South America, and Europe, glaciers reached their greatest development during the glacial ages. This development involves the close genetic relationship among valley glaciers, piedmont glaciers, and ice sheets.

VALLEY GLACIERS, PIEDMONT GLACIERS, AND ICE SHEETS

Glaciers exist today wherever the necessary combination of adequate snowfall and low temperature (especially low summer temperature)

[5] Paschinger 1923.



Foresta H. Wood

FIG. 6. Three small plateau ice caps, St. Elias Range, Yukon Territory, Canada. The plateaus alternate with valleys containing valley glaciers with massive lateral and end moraines of recent date. In the foreground is the terminal zone of the large Wolf Creek Glacier, blanketed with ablation moraine.

exists. In general this combination is found in high latitudes and at high altitudes. The kind of glacier that develops in a given region depends partly on the configuration of the ground, but even more on the local ratio of snowfall to wastage. If the ground is more or less flat and plateaulike the accumulating snow (provided it is not blown away by the wind before it becomes névé) takes the broad form of an *ice sheet*, or *ice cap* as it is sometimes called (Fig. 6). Plateau ice sheets, some of them very small because the plateaus beneath them are small, are known in northern Norway, in Spitsbergen, on some of the eastern islands of the Canadian Arctic archipelago, in the Alaska-Yukon region, and in the southern Andes.

On the other hand, ground that has been dissected by streams so as to form systems of valleys prior to the accumulation of glaciers is much more common than undissected plateaus. On ground like this, fallen snow accumulates and persists mainly in and near the heads of the valleys and forms valley glaciers, whose pattern is therefore determined by the pattern of the valleys. At this point the ratio of snowfall to wastage plays an important part. If wastage balances snowfall at any point before the valleys have become completely filled with ice, the glaciers remain valley glaciers. This is the common condition under the climates of

Bradford Washburn

FIG. 7. Columbia Glacier, a piedmont glacier, Prince William Sound, Alaska. Calving into tide water, the piedmont lobe is shrinking rapidly.

today; the glaciers of the Alps are the most widely known and most persistently studied example.

But let the valley glaciers reach the base of the mountains, so that their terminal parts spread out on the lowlands beyond, before equilibrium between snowfall and wastage is reached, and the valley glaciers become piedmont glaciers (Fig. 7), such as are found today on the Alaskan, Greenland, and Antarctic coasts. A very slight warming of the climate would destroy the broad piedmont termini by ablation and convert the piedmont glaciers into valley glaciers. On the other hand a reduction of temperature would increase the persistence of snow through the summer and also the volume of ice (including the piedmont ice) and, if continued, would lead to the growth of an ice sheet without the necessity of beginning with an undissected plateau. The ice sheet would simply bury the mountains.[6]

[6] An extensive ice sheet formed in this way is called by some, especially in Scandinavia, *inland ice.*

An ideal example[7] is presented by a high mountain range trending north-south, standing above broad lowlands, and receiving abundant snowfall precipitated from moist westerly winds (Fig. 8). With sufficiently low temperature, especially in summer, small glaciers would form on both flanks of the mountains. They would lengthen, forming a system of valley glaciers, until they reached the lowlands. Here they would spread out as piedmont glaciers, and adjacent piedmonts would

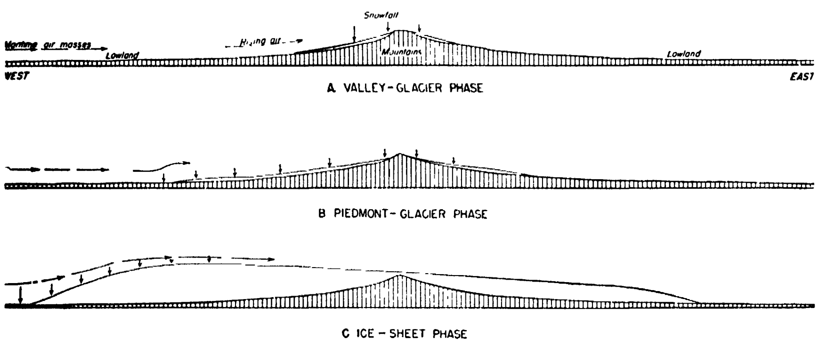


FIG. 8. Vertical sections illustrating ideally the development of an ice sheet in mountains surrounded by a lowland. (Cf. Figs. 49, 50, and 61.)

coalesce to form continuous ice masses along both windward and leeward bases of the mountains. The locus of maximum snow accumulation, at first confined to the mountains, would be spread through a zone of increasing width owing to the effect of the expanding piedmonts, which, being both higher and colder than the same areas before glaciation, would induce increased snowfall upon them. It has been calculated[8] that, whereas about 20 per cent of the solar radiation received per unit area by a land surface not covered by ice or snow is reflected back into space, about 80 per cent of that received by a glacier- or snow-covered area is lost by reflection, and that the lost portion would be sufficient, in temperate latitudes, to melt more than 30 feet of ice over the unit area annually. Not only would this enormous loss of heat reduce ablation; by decreasing the temperature of the highland it would also induce increased precipitation. The resulting thickening of the piedmonts would make an increasingly high topographic obstacle for the winds, which, in addition to the factor of low temperature, would make for increased precipitation.

[7] Flint 1943.
[8] C. E. P. Brooks 1928, p. 128.

Obviously the windward piedmont would receive more precipitation than the leeward piedmont, and therefore glacier expansion would be more rapid on the windward side. Unless the general climatic conditions changed, the thickening and spreading of the piedmonts would continue until the ice overtopped the mountains and became a true ice sheet. Although the ice sheet might have a considerable extent to leeward, its area of most active nourishment would be its windward flank where it faced its principal source of moisture. Earlier, when the ice consisted only of valley glaciers, the ice would have flowed away in opposite directions from the crest of the mountains. But, once buried, the mountains would no longer exert this influence on the direction of flow of the ice. Instead, the center of radial outflow of the ice sheet would have shifted to some area to windward of the mountains, where under the new conditions the snowfall was greatest.

As the ice sheet thickened, the level of maximum snowfall would be lowered somewhat. In consequence, nourishment by snowfall, at first distributed without too much inequality over the entire surface of the ice, would gradually be concentrated toward the windward margin at the expense of the central and leeward parts. This would have made the ice sheet thickest in its windward part. Its maximum thickness, of course, would have been that thickness at which equilibrium was reached among the three factors of rate of accumulation, rate of flow, and rate of wastage. But back of these factors lies the fact that, the higher the ice sheet was built above the level of maximum snowfall, the less would be the snowfall upon its highest part. Diminishing accumulation above the critical level of maximum snowfall, therefore, would have set a limit to thickness, a limit which in middle latitudes with abundant precipitation would probably have been in the neighborhood of 10,000 feet. This figure, for what it may be worth, is of the same order of magnitude as the thickness estimates arrived at by considering the amount of crustal warping induced by the building and later disappearance of the former Scandinavian Ice Sheet (Chapter 19).

To be sure, the development just sketched is partly theoretical. But the great ice sheets of former times were not confined to plateaus, and a good deal of direct evidence points to their having developed in this way. Furthermore in existing glaciers we can observe gradations from one main type to another. Gradation from valley glaciers to piedmont glaciers is found today wherever ice tongues flow from steep narrow valleys on to surfaces of gentle slope. Gradation from piedmont glaciers to ice sheets, though much less common, is probably represented by the great glacier system between Mt. St. Elias and Mt. Logan in coastal Alaska, as well as in western Spitsbergen.

GREENLAND ICE SHEET[9]

Area, Form, and Thickness

The Greenland Ice Sheet (Fig. 9) occupies all of Greenland except a narrow belt of country around the coast. It is nearly 1500 miles long and

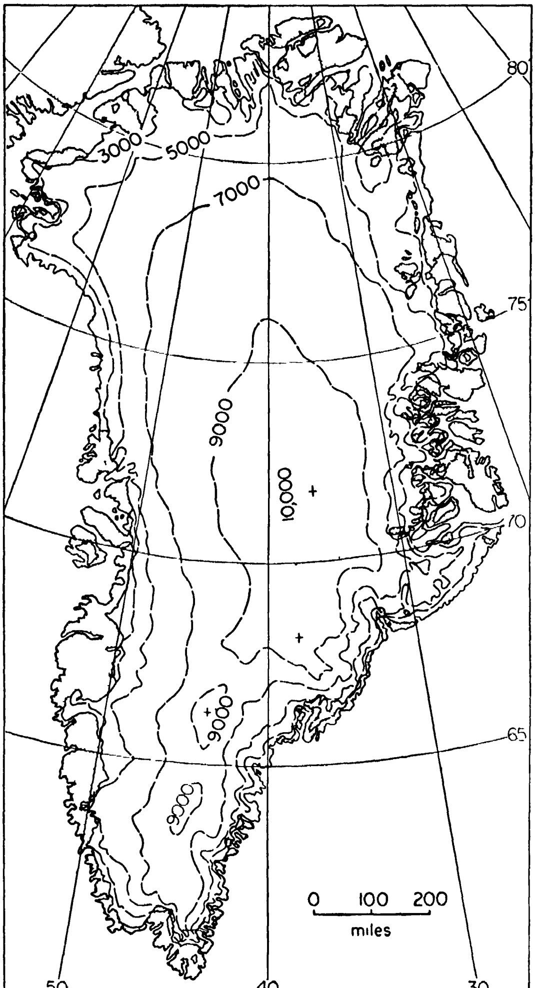


Fig. 9. Sketch map of Greenland. Ice-free areas are shaded. The broken lines are form lines (generalized contours) on the ice sheet and are not accurate in detail. Altitudes are given in feet. Crosses mark the approximate summits of known domes on the ice sheet. (Compiled from various sources, including unpublished data supplied by F. A. Wade and H. G. Dorsey, Jr.)

[9] Good general references are Demorest 1943; Kayser 1928; Koch 1923; Nordenskjöld

200 to 500 miles wide, and has an area of about 637,000[10] square miles. In addition to the main ice sheet, Greenland has hundreds of independent glaciers occupying the coastal belt in both East and West Greenland. These glaciers include both valley glaciers and plateau ice caps, some of which are of considerable size. Some of these may be relics of the ice

TABLE 1. APPROXIMATE AREAS OF EXISTING GLACIERS BY GROUPS
With slight changes, figures are from Thorarinsson (1940), modified after Hess (1933).

	km.²	mi.²	km.²	mi.²	km.²	mi.²
Alps	5,000	1,930				
Pyrenees	40	15				
Scandinavia	6,280	2,416				
Total: Continental Europe			*11,320*	*4,370*		
Caucasus	2,000	772				
Central Asia and Siberia	110,000	42,471				
Total: Continental Asia			*112,000*	*43,243*		
Africa	20	8				
Total: Continental Eurasia					*123,320*	*47,613*
New Guinea	15	6				
New Zealand	1,000	386				
Total: Pacific Islands					*1,015*	*392*
Continental North America	80,000	30,888				
Continental South America	25,000	9,652				
Total: Continental Americas					*105,000*	*40,540*
Sub-Antarctic Islands	3,000	1,158				
Antarctic Continent	13,000,000	5,019,300				
Total: South Polar Region					*13,003,000*	*5,020,458*
Iceland and Jan Mayen	12,600	4,864				
Spitsbergen (including North East Land)	58,000	22,393				
Franz Joseph Land	17,000	6,563				
Novaya Zemlya	15,000	5,791				
Severnaya Zemlya	15,100	5,830				
Greenland	1,650,000	637,065				
Canadian Arctic Islands	100,000	38,610				
Total: North Polar Region					*1,867,700*	*721,116*
Grand total: Existing glaciers of the world					*15,100,035*	*5,830,119*

Percentage of land area of world (taken as 57,800,000 square miles) presently covered by ice = 10.1%.

sheet, left over from the time when it was more extensive than now. Others probably have grown up *de novo* since the ice sheet uncovered the areas where they now occur. Over its broad interior the slopes are very gentle (5 to 50 ft. per mile), but close to its margin they increase in places to several hundred or even 1000 feet per mile (Fig. 10). The

[10] This and other area figures are shown in Table 1.

margin is very irregular in plan owing to the presence of high rugged coastal mountains, through whose valleys the ice escapes to the sea as tonguelike *outlet glaciers.* Intricately crevassed, these outlet tongues reach the sea on fronts commonly several miles in width, and their high cliff-like termini calve icebergs into the deep water. Humboldt Glacier (at lat. 80° on the west coast), the largest of these outlets, is 60 miles wide at its terminus. The vast central area of the ice sheet, though its slopes are gentle, nevertheless rises to form at least three[11] broad domes, the

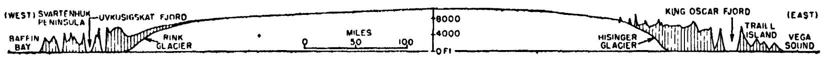


FIG. 10. Profile across the Greenland Ice Sheet at latitude 72°, drawn so as to include two representative outlet glaciers, Rink Glacier on the west and Hisinger Glacier on the east. Vertical scale exaggerated about 21 times. (Compiled from various sources.)

high points of the ice surface, each lying somewhat east of the geographic long axis of the ice sheet as a whole (Fig. 9). The highest of them reaches an altitude that may exceed 11,000 feet.

The thickness of the ice and the nature of the floor beneath it are almost wholly matters of conjecture. Except in the marginal zone no rock projects through the ice. This has given rise to a belief that the ice sheet is very thick, but the inference is by no means necessarily true. An attempt at measurement by seismic methods was made by the Wegener Expeditions of 1929 and 1930–1931, which made a traverse consisting of 14 soundings near latitude 71°. All but one of these were located within 80 miles of the western margin of the ice sheet; the exception was near the geographic center. In analyzing the results,[12] Demorest has shown[13] that the soundings are so placed, and so few in number, that they do not afford a basis for an estimate of the position and nature of the floor beneath the ice sheet. The ice-thickness figures obtained ranged, near the margin, between 150 feet and 6100 feet. At the one central station, the result obtained was unreliable. The most that can be said is that within the limited area of the soundings the underlying surface appears to be one of great relief. Demorest showed further that there appears to be no factual basis for the statement[14] that the floor beneath the ice sheet is saucerlike, having been depressed beneath the weight of the ice.

Because knowledge of the subglacial floor is all but nonexistent, the origin of the ice domes is an unsolved problem. Two principal modes

[11] Lindsay 1935, pp. 406, 408; Loewe 1935, p. 347.
[12] Detailed by Brockamp, Sorge, and Wolcken 1933.
[13] Demorest 1943, p. 383.
[14] Cf. Daly 1934, pp. 9–11.

of origin may be considered. According to the first, clearly expressed by Koch,[15] the domes occupy two conspicuous highland areas and are the topographic expression of those high areas, thinly covered with ice. The other, implied by Wager[16] among others, is that the domes consist of exceptionally thick ice. Such domes must result from exceptionally great accumulation at the sites of the domes, now or in the very recent past, as compared with other parts of the ice sheet.

The data given in the following paragraphs on the distribution of accumulation on the ice sheet do not bear out the second mode of origin. Furthermore, as Demorest pointed out, present meteorologic conditions on Greenland should lead to the accumulation of a single elongate dome rather than two or more separate ones. For these reasons Demorest[17] favored the first mode of origin.

Regimen

Nourishment

The "glacial-anticyclone" hypothesis. In repeated publications beginning in 1911 Hobbs[18] advocated the view that Greenland is nourished by an "anticyclone" set up by the presence of the ice cap. According to this view cooled and therefore heavy air flows downward and outward along the ice surface to the sea in "strokes" or "pulses" caused by cooling of the surface air by radiation of heat from it. The outflowing air, which moves with great force, is replaced by air drawn down toward the ice sheet from aloft. As it descends the new air is warmed adiabatically (by compression). The warmth raises the air temperature at the surface of the ice so much that the air stops flowing downward and outward and the stroke comes to an end. Meanwhile as the descending air is cooled by contact with the ice surface some of its contained moisture condenses and in the form of rime or hoarfrost adds substance to the ice sheet. The cooling process eventually reaches a point at which another stroke occurs, and the minute particles of condensed moisture are blown outward to lodge in the marginal zone of the ice sheet, which is conceived of as being nourished largely by this mechanism rather than by cyclonic and orographic snowfall.

This ingenious concept is picturesquely described by its author[19] in these terms:

[15] Koch 1928, pp. 434–446.

[16] Wager 1933, p. 154.

[17] Demorest 1943, p. 381.

[18] Cf. Hobbs 1926. Since the completion of the present work, an extensive study and adverse criticism of that hypothesis has been completed by F. E. Matthes for publication in the Transactions of the American Geophysical Union.

[19] Hobbs 1926, p. 1.

> Inland-ice . . . appears to be regularly fed from *high-level* currents through the operation of a refrigerating air engine of which the ice mass and its atmospheric cover are the essential parts.
>
> Through the rhythmic action of this engine the congealed moisture derived from the ocean surface within moderate or low latitudes and carried to the polar region in the high-level cirrus clouds, is pulled down to the surface of the glacier in the eye of a great glacial anticyclone which is centered above it. During their descent from the high levels the ice grains of the cloud are melted and vaporized by adiabatic warming, and on reaching the cold surface layer of air next the ice, are quickly frozen to form flakes of fresh snow. The progressive warming of the air adiabatically both during its descent to the central area of the ice mass and on the further slide outward to the peripheral portions, gradually damps and eventually stops the sliding centrifugal motion of the surface air-layer. Thus the engine comes to rest or, as we may say, has reached the end of its stroke. The great calm which ensues allows heat to be again slowly abstracted from the surface layer of air, thereby lowering its temperature and raising its density until gravity again starts the engine, which now acquires the steadily accelerating velocity characteristic of bodies sliding on inclined planes. The tempest which is eventually engendered is succeeded by a rapid rise of air temperature, a fall of fresh snow, and another stopping of the engine.
>
> The fierce violence of the surface air currents when at their maximum, and the fall of the snow for the most part as the engine is slowing down, together make of this glacial anticyclone a gigantic snow broom. The snow deposited as it were between strokes of the engine is by the next sweep of the broom brushed largely clear from all central portions of the glacier, and the sweepings are deposited near and about the margins of the mass.

In commenting on this idea we may begin by remarking that the outblowing wind described is not properly an anticyclone. That term commonly refers to a system of air flow that may be very deep and that is caused chiefly by differences in pressure at the same altitude.[20] The Greenland wind described is a *katabatic* ("down-flowing") *wind,* shallow, controlled by gravity, pouring off the ice surface like a sheet of water, and independent of the distribution of pressure at any one altitude.

It appears to be generally recognized that (1) centripetal flow of upper air with respect to the ice sheet, as observed from the movement of cirrus clouds, (2) katabatic winds, and (3) supply of moisture, chiefly as rime, in small quantity to the central area of the ice sheet, do in fact occur. On the other hand: (1) calms appear to be far less common than

[20] Cf. Napier Shaw 1927, p. 29.

the glacial-anticyclone hypothesis requires; (2) the domed form of the ice sheet prevents the accumulation of much cooled air upon it; (3) the outbreaks of cold air from the ice sheet occur at times when favorable pressure gradients exist because of temporary conditions in the general atmospheric circulation.[21] These considerations are unfavorable to the glacial-anticyclone concept. An alternative hypothesis appears to explain the facts much more satisfactorily.

ALTERNATIVE HYPOTHESIS. The data assembled by Demorest[22] from observations by Koch and Wegener, de Quervain, and the Wegener Expedition of 1930–1931 show clearly that in northern as well as in southern Greenland accumulation is greatest, not at the crest of the ice sheet, but in elongate belts parallel with the crest and at some distance from it. They show, further, that accumulation is greater on the western side of the crest than on the eastern. This distribution of accumulation does not fit the glacial-anticyclone concept. On the other hand, it is what would be expected if the ice sheet were nourished chiefly by snowfall from cyclonic storms approaching it from the southwest and occasionally moving across it. On this hypothesis low-pressure areas crossing or moving along the flanks of the high cold barrier precipitate moisture on its windward slope just as they do on the comparable barriers of the Sierra Nevada and the Rocky Mountains. Shallow cyclones encountering the ice sheet are unable to pass over it. Usually they are deflected north along the west coast of Greenland, causing precipitation on the marginal zone of the adjacent ice sheet. Deep cyclones, extending into the upper air, pass across the ice sheet, flattening somewhat in the process, to be sure. Whether they pass across the ice sheet or are deflected by it, these cyclones apparently furnish the chief nourishment to the broad west flank of the ice.

The smaller precipitation on the east flank, which diminishes greatly toward the north, appears to come from air circulating counterclockwise around the Iceland barometric depression and being lifted somewhat as it encounters the ice sheet. Finally, in the vicinity of the crest of the ice, accumulation may include a small proportion of hoarfrost and rime.

With the facts available at present these three sources of moisture appear to account satisfactorily for the nourishment of the ice sheet. But the matter will never be fully cleared up until more thorough meteorological surveys of Greenland have been made.

Before leaving the subject of the wind system and the type of precipitation associated with the Greenland Ice Sheet, we must note a widespread misconception regarding the outblowing winds. The katabatic winds

[21] H. G. Dorsey, Jr., *unpublished.*

[22] Demorest 1943, p. 378.

of Greenland are confined to the ice sheet and its immediate vicinity. Beyond the ice sheet they funnel down the radial, channel-like fiords with great force, but once they reach the coast their force dissipates. This is natural because of their dependence on gravity for their motion. Partly because the term *anticyclone* has been improperly used to describe these winds, there has been a general tendency to think of glacial-anticyclonic winds as blowing off the former great ice sheets of North America and Europe and affecting very broad belts of territory beyond them. That strong cold winds frequently moved from various sectors of these ice sheets across the adjacent territory is not doubted, but these winds apparently were normal outbursts of cold air such as now occur each winter in our northern continental interiors. Because of the presence of the ice sheets they were generally stronger and originated farther south than those that characterize the atmospheric circulation today. The former ice sheets, like the present Greenland Ice Sheet, were crowned by persistent wedges of cold polar air characterized by high-pressure, anticyclonic conditions. Yet it is probable that, between the times of outburst of such air into the surrounding region, cyclonic air masses were able to invade the marginal parts of the ice sheets through considerable distances from their peripheries.

Distribution of precipitation. On both east and west coasts precipitation diminishes very rapidly from south to north. The figures in Table 2 illustrate this point.

Table 2. Temperature and Precipitation at Coastal Stations in Greenland

Station	*Latitude* (N)	*Altitude* (feet)	*Mean Annual Temperature* (°F)	*Mean Annual Precipitation* (inches)
		WEST COAST		
Ivigtut	61°12′	82	33	45
Godthaab	64°11′	66	29	24
Godhavn	69°15′	26	23	15
Upernavik	72°47′	59	17	9
Thule	76°34′	23	9	3
Ft. Conger[23]	81°47′	(approx.)20	−4	4
		EAST COAST		
Angmagssalik	65°37′	95	29	34
Scoresbysund	70°29′	230	17	12
Myggbukta	73°29′	10	13	3
Danmarkshavn	76°46′	20	9	6

[23] East coast of Ellesmere Island, 26 miles west of the west coast of Greenland.

Probably the gentler northern slopes of the ice sheet (especially in its marginal part) (Fig. 9) reflect this sharp northward decrease in precipitation. There is not enough accumulation in the north to build up the flanks of the ice sheet to steeper slopes. In consequence the maintenance of the glacier in its northern part is dependent on transfer of ice by flow from areas farther south, small local accumulation, and greatly reduced ablation because of low temperatures and primarily north-facing slopes — that is, slopes facing away from the Sun. These conditions in northern Greenland seem to be as old as the ice sheet itself, for although the rest of coastal Greenland, now ice free, apparently was overwhelmed by the ice sheet during the glacial ages, the northern part of Peary Land is reported to show no evidence of ever having been covered by the ice sheet. This is not surprising, inasmuch as a thickening of the windward part of the ice sheet should, in theory, only increase the starvation of its cold most northerly part.

Movement and Wastage

As no systematic comparative measurements of the amount of accumulation on various parts of the ice sheet have been made, we can still only estimate very roughly the regimen of this vast glacial unit. Precipitation on the surface of the ice sheet apparently does not exceed 15 to 20 inches of water annually. On the other hand, ablation amounting to 90 inches annually has been measured near the west coast. The discrepancy between these two figures is accounted for in two ways. First, the ablation area is far smaller than the area of accumulation. (The névé line varies considerably in altitude, but it lies everywhere near the edge of the ice sheet.) Also glacial flow from beneath the central part of the ice sheet brings ice to the area of ablation. The rate of movement is negligible beneath the central mass but is considerable in some of the outlet glaciers through which the discharge reaches the sea.

In addition to the measured ablation, the ice sheet loses substance by calving of bergs at the termini of the many outlet glaciers that end in the deep water of fiords and bays. Smith estimated that West Greenland outlet glaciers discharge 7 to 10 cubic (nautical) miles of ice in the form of bergs into the sea each year.[24] The iceberg output of the East Greenland glaciers has not been calculated, but it is believed to be about the same as the figure for West Greenland.

It is doubtful whether any considerable part of the Greenland Ice Sheet has been in a state of even approximate equilibrium during the twentieth century or perhaps even during the latter part of the nineteenth.

[24] E. H. Smith 1931, pp. 74, 189.

Current observations indicate that most if not all outlets have suffered net losses during this period. Comparative observations suggest that the terminus of the large Pasterze Glacier tributary to Tyroler Fjord on the east coast at latitude 74°40′ receded about 4 miles during the 68-year period 1869–1937.

ORIGIN

The origin of the Greenland Ice Sheet has not received much study. It appears probable that the ice began as valley glaciers and small plateau ice caps in the highlands and grew up by gradual coalescence of these units. C. F. Brooks[25] and Kayser[26] stressed the coastal mountains as the sites of the first glaciers. Koch[27] stressed the highlands, which he believed directly underlie the domes at the crest of the ice sheet and which, if present, must be nearly as high as the highest coastal mountains. If the buried highlands are a reality, then probably both views are essentially close to the truth, and the highest parts of Greenland, whatever their character in detail, were the places of origin of the glaciers that later became the Greenland Ice Sheet.

In late preglacial time, before the ice sheet with its centrifugal katabatic winds had come into existence, cyclonic air masses probably crossed Greenland more frequently than they do today. Also, because atmospheric pressure over the polar area was presumably then somewhat less strong than now, some of these cyclones probably followed more northerly paths than at present. As a result the conditions were favorable for the distribution of moisture to all the highlands of Greenland.

With the reduction of temperature that began the first glacial age these highlands should have developed small valley glaciers and plateau ice caps. These expanded and coalesced in the intervening lowlands to form piedmont glaciers and later a single continuous ice sheet. The loci of maximum accumulation were thereby shifted from the highlands to the ice sheet itself and may have occupied many different positions on the ice sheet during its existence. Centrifugal, katabatic winds developed and reinforced the high cold ice sheet itself as a fender against traveling cyclones. During the height of at least one glacial age the ice sheet was considerably thicker and more extensive (so far as the depths of water off the coast would permit) than it is now.[28] It may be that at that time no cyclones could cross Greenland, but even then the flanks of the ice sheet were nourished by snowfall from depressions moving north-

[25] C. F. Brooks 1923, p. 452.
[26] Kayser 1928, p. 366.
[27] Koch 1928, pp. 434–446; see also Demorest 1943, p. 383.
[28] Cf. Bretz 1935, p. 189.

ward along the west coast and, to a lesser degree, by the moist air that reached the east coast from the southeast. Air from that source may have been less moist than it is today owing to the fact that the Iceland low then lay farther south than now, separated from the east coast of Greenland by a broad expanse of pack ice.

If this reconstruction approximates what actually happened, then future shrinkage of the Greenland Ice Sheet should be accompanied by the persistence of local glaciers in the coastal mountains and of residual ice caps on whatever highlands may today lie close beneath the central parts of the ice.

Demorest[29] attached significance to the fact that, except for outlet glaciers, the low-level ice around the margins of Greenland has vanished, while the high-level ice, dammed up, as it were, in the interior of Greenland back of the coastal mountains, still persists and still receives nourishment. At the height of the most recent glacial age the low-level ice on the east coast certainly extended to some distance offshore, while on the west coast it may have extended far enough west, partly aground and partly afloat, to have been coalescent with the glacier ice of Baffin Island.

After the rise in temperature that started the last great deglaciation, the disappearance of the western low-level ice was hastened, according to Demorest, by the emergence of the higher coastal islands and peninsulas as nunataks. The bare-rock surfaces of these projecting masses reflected solar heat, and the sun-heated rock set up convection currents of warmed air, which, moving across the glacier surface, promoted melting and evaporation. Melting was promoted further by summer rains precipitated from the same air masses, reaching Greenland from the southwest, that had nourished the ice sheet with snow at an earlier time when regional temperatures were lower.

In contrast the high-level ice is still free of nunataks except at its extreme margins, and it is neither too cold nor too broad to receive appreciable increments of nourishment. In contrast the conditions on the Antarctic Continent, described in the following section, appear to be the reverse of the conditions in Greenland.

ANTARCTIC ICE SHEET

Area, Form, and Thickness

Most of the Antarctic Continent lies within the parallel of 70° south. The area of the continent is not known with accuracy, but it is believed to be slightly more than 5,000,000 square miles. Of this area all but a few

[29] Demorest 1943, p. 395.

hundred square miles is covered by the Antarctic Ice Sheet. This greatest of all ice sheets is about eight times more extensive than the Greenland Ice Sheet and has more than half again the area of the United States. It is unsymmetrical, being highest not at the South Pole but along a line roughly parallel with the Pacific coast and lying between the Pole and

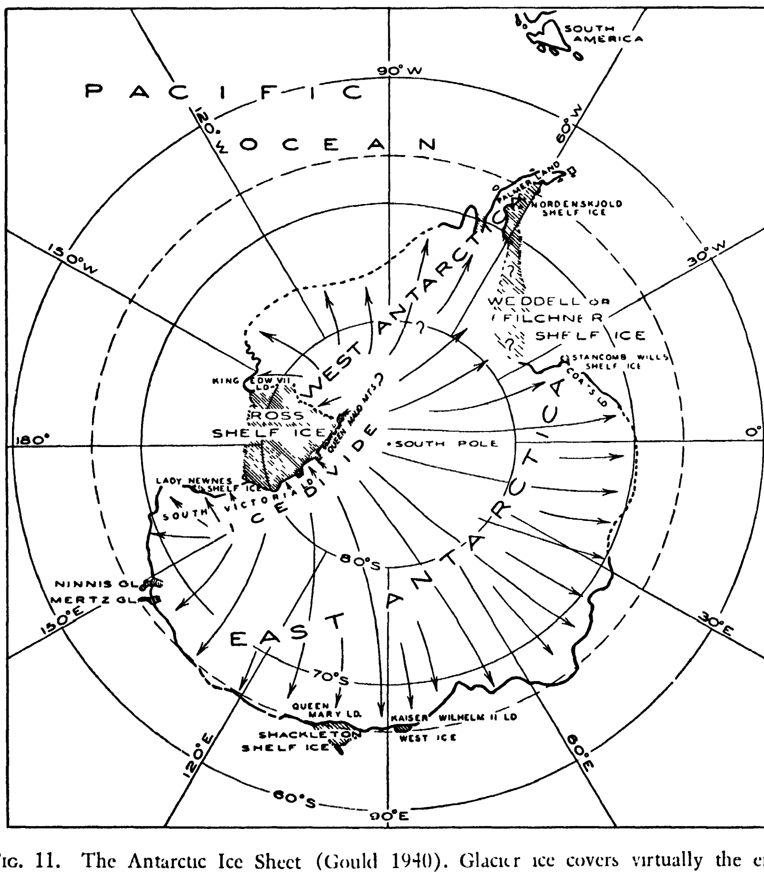


Fig. 11. The Antarctic Ice Sheet (Gould 1940). Glacier ice covers virtually the entire Antarctic Continent.

that coast. This high crest, about 10,000 feet above sealevel, in turn lies parallel with and inland from a great mountainous highland reaching maximum altitudes of more than 12,000 feet and having a length of more than 1000 miles. Poleward from the crest the ice sheet slopes gently away across the south polar region to the coast, a distance along various radii of 700 to 1400 miles. The ice enters the sea along wide fronts unbroken by nunataks or confining highlands. Westward from the

crest the ice sheet flows through gaps in the great mountainous highland and thus subdivides into outlet glaciers.

The average altitude of the Antarctic Continent is high. Debenham[30] described this land mass as a nearly continuous plateau with an average altitude of 6000 feet and a maximum altitude of nearly 10,000 feet.

Probably the ice is relatively thin and fails to mask entirely the major irregularities of the rock surface beneath it. It is unlikely that the ice thickness exceeds 2000 feet except very locally; probably its average thickness is considerably less.[31] Most observers are now agreed, although the evidence is still extremely meager, that the form of the ice sheet, far inland as well as along the coast, is controlled to a considerable degree by the irregularities of the underlying terrain.

A conspicuous feature of the margin of the Antarctic Ice Sheet is the fact that in many coastal sectors the ice is afloat. In these places the glacier has pushed out into deep water until the ice has been buoyed up. In some sectors at any rate the floating ice is at least 1000 feet thick. The largest and most conspicuous of these *shelf ice* units is the Ross Shelf Ice[32] which half fills the vast embayment known as the Ross Sea. The area of this unit is about 200,000 square miles, almost as large as the State of Texas. Its seaward face is a cliff that reaches a height of 200 feet above sealevel, and it calves off gigantic bergs, some of which are 30 miles in longest diameter. It is probable that during the heights of the glacial ages many coastal sectors in both North America and Eurasia were marked by vast fringes of similar shelf ice. One of the most notable of these was the 1000-mile sector along the outer edge of the continental shelf of North America between Long Island and Newfoundland.

Regimen

If little is known about the form and extent of the Antarctic Ice Sheet, even less is known about its regimen. Precipitation is slight, but wastage is slight also. In consequence the regimen is nearly in balance. Observations made over a period of many decades indicate that the ice sheet is shrinking, but that it is shrinking very slowly.

At the Little America station, at the outer margin of the Ross Shelf Ice, the annual precipitation (entirely snowfall) is equivalent to 7.5 inches of water. C. E. P. Brooks[33] estimated that the precipitation nowhere

[30] Debenham 1943.

[31] David 1914, p. 615; Debenham 1943; Gould 1940. David argued that in the vicinity of the magnetic pole (lat. 70–73°S; long. 148–156°E) this is suggested by the great strength of the local variations in magnetic declination. If the ice blanket were many thousands of feet thick it would make the variations weaker and more generalized.

[32] Gould 1935.

[33] C. E. P. Brooks 1930, p. 188.

exceeds 10 inches, and that in most places it is less than 5 inches. Since in no month does the mean temperature rise above the freezing point, ablation is slight at all times, and the small precipitation results in effective accumulation of ice. The névé line lies everywhere at or very close to sealevel.

The winds in the coastal sectors in which observations have been made are prevailingly southeast. According to Brooks they are katabatic winds like those encountered in Greenland, outflows of heavy cold air sliding down the radial slopes of the ice sheet to the sea and, in the process, deflected to the left (westward) by rotation of the Earth. Maritime air masses from over the seas surrounding the Antarctic Continent reach the coastal areas at least and contribute snowfall to the marginal parts of the ice sheet (Plate 1). Accumulation of ice is somewhat more than balanced by slow seaward flow of the ice sheet with marginal loss by calving, a process that probably accounts for ice losses exceeding those caused by melting and evaporation combined. Not only is the ice sheet shrinking at present; the geologic evidence indicates clearly that it has shrunk notably within the very recent geologic past. The observations of David and Priestley,[34] Gould,[35] and others showed that the ice sheet was formerly at least 1000 feet thicker than it is today, that some of the outlet glaciers were 2000 to 3000 feet thicker than now, and that the Ross Shelf Ice was so thick that it was aground instead of being afloat. The area of the ice sheet, correspondingly, was greater.

The deglaciation that has occurred since the time when the ice sheet was thicker and more extensive can not be the result of increased ablation, because, as already pointed out, ablation even today is a minor if not negligible factor — the temperature even along the coast rarely exceeds 0° C. Shrinkage of the ice sheet results apparently from the very low precipitation, which is in part the result of the extreme cold induced by the presence of the ice sheet itself. The great central area of the ice sheet is so undernourished that it is literally starving. It is approaching a condition in which it is affected by neither accumulation nor ablation, and in consequence it is gradually thinning and spreading out.

If this condition should continue indefinitely, it seems probable that the high-level ice in the interior would thin to the point of stagnation, cutting off the supply to the Ross Shelf Ice and other parts of the low-level ice of the coastal fringe. But the low-level glaciers should be able to maintain themselves by snowfall upon their own surfaces — snowfall largely denied to the high-level ice because of the extreme cold and the very low atmospheric humidity that prevail over it. This relationship

[34] David and Priestley 1914, pp. 46, 290.
[35] Gould 1940, pp. 863–864; see also Hobbs 1910*a*.

stands in sharp contrast with the conditions in Greenland, where the once extensive low-level ice has already almost entirely disappeared, leaving the high-level ice sheet in a slowly shrinking but apparently healthy condition.

It was long ago deduced by Scott that, if the climate of the Antarctic region became warmer, the Antarctic Ice Sheet would not shrink; on the contrary it would expand.[36] This deduction applied to the central body of the ice, not to the ice of Graham Land (Palmer Peninsula) which reaches beyond the Antarctic Circle to latitude 65°. The warmer air could hold more moisture; evaporation would increase, precipitation would be more abundant, and the central part of the ice sheet would thicken. On the other hand he thought the colder climates that accompanied the several glacial maxima in the northern hemisphere could only have increased the present-day conditions of starvation in the main body of the Antarctic ice.

In an interesting discussion of this paradox Gould[37] emphasized that the islands fringing the Antarctic Continent show evidence of stronger former glaciation than the Antarctic mainland, although their climates are notably warmer than the mainland climate. On them both precipitation and ablation are much greater than on the ice sheet itself. It can hardly be doubted that the strong glaciation of these islands occurred during times of colder climate, coinciding with the heights of the glacial ages widely recorded in the northern hemisphere. On the other hand, if Scott's deduction be accepted, then the times when the Antarctic Ice Sheet was thickest and most widely expanded would seem to have coincided with the earlier parts, rather than the maxima, of the glacial ages, when worldwide climatic chilling had begun but had not yet reached the starvation point for the Antarctic region. Essentially this idea was expressed by Simpson.[38] Unfortunately he tried to apply it too widely, and it did not gain the following it deserves.

However, the view has been put forward that a decrease in temperature would reduce the rate of flow of the central body of the ice sheet, which would therefore thicken during a glacial age, synchronously with glaciers in lower latitudes.[39]

The Antarctic glaciers appear to be shrinking today, just like the glaciers in all other parts of the world. To be sure, the baseline for measurement is short, for Antarctic observations not only are extremely scanty but extend back only to 1843 when Ross first charted the Ross

[36] For a detailed analysis of this view, see Meinardus 1925–1928.
[37] Gould 1940, p. 867.
[38] G. C. Simpson 1934, p. 431.
[39] Nolke 1932.

Shelf Ice. Shrinkage of the glaciers in the Ross Sea region (lat. 80°) is the result of starvation, whereas shrinkage in the Graham Land region (lat. 65°–70°) is the result of ablation despite a fair amount of nourishment.

Very likely glaciers existed in the Antarctic long before the vast ice sheet we know today was formed. Indeed it is quite possible that the highest lands of the Antarctic Continent were cold enough to support local glaciers before the Pleistocene epoch, as it is commonly defined in the northern hemisphere, had officially begun. Gould[40] suggested that the Antarctic Ice Sheet came into existence as the result of the growth and coalescence of many local glaciers on the high mountains and plateaus of the south polar region. This concept is identical with the opinions quoted on the origin of the ice sheet in Greenland.

The vast former ice sheets of North America, Europe, and western Siberia were climatically more like the Greenland Ice Sheet than like its larger Antarctic counterpart. None of them were in regions of extreme cold; all of them lay in the paths of traveling cyclonic air masses; they tended to receive abundant precipitation and to be affected, at times at least, by strong ablation. If analogies with present conditions are to be pressed, we must look to Greenland rather than to Antarctica for them.

ICELAND AND JAN MAYEN ISLAND

The two great ice sheets of the Antarctic Continent and Greenland together constitute about 99 per cent of the combined area of all existing glaciers. The remaining 1 per cent include many small ice sheets and valley glaciers. A survey of their distribution (Plate 2) is essential to an understanding of the distribution of the expanded glaciers of the recent geologic past.

ICELAND[41]

Approximately one-eighth (about 4900 square miles) of the area of Iceland is occupied by glaciers, of which one, *Vatnajökull* (*jokull* = glacier), is larger than the others combined. All are plateau ice caps, and the larger ones have outlet glaciers which descend the valleys leading from the plateaus to the sea. None of these outlet glaciers now reaches sealevel. The ice caps are not true domes; they conform to the relief of the underlying plateau (which reaches extreme altitudes of more than 6000 feet), and the largest probably does not exceed 750 feet in thickness at any point.

[40] Gould 1940, p. 868.
[41] General reference: Bárðarson 1934.

Along the south coast, where temperatures and precipitation are high, the névé line is at about 3600 feet. In the interior, where although temperatures are lower precipitation is sharply decreased, the névé line stands at more than 4500 feet. In the northwest, where despite reduced precipitation temperatures are very much reduced, the névé line descends to about 2000 feet. Vatnajökull, the largest ice cap, is maintained by snowfall from moist air masses derived from the Gulf Stream, closely adjacent to the south. The maximum measured mean annual precipitation on Vatnajökull amounts to 170 inches, but this high figure is offset by ablation equivalent, at the southern margin of the glacier, to 40 feet of ice annually.

The ice caps of Iceland in general and Vatnajökull in particular thus have an exceedingly high "metabolic rate," owing to the maritime climate of the island. Their intake of nourishment is high, but their losses are high also. As might be expected, their rates of flow are comparatively great. In all these respects they differ markedly from the Antarctic Ice Sheet and the northern part of the Greenland Ice Sheet, where extreme cold inhibits nourishment and ablation and where in consequence rates of flow are very small.

Jan Mayen Island

Jan Mayen, a small island in the Greenland Sea about 300 miles northeast of Iceland, is dominated by the Beerenberg, a conical volcanic peak with an altitude of 7680 feet. This peak is the site of six long glaciers and many shorter ones, which spread outward on its flanks. The largest heads in the volcanic crater itself, and some of those on the north slopes reach sealevel, where they calve into deep water. As the valleys on the flanks of the cone are shallow at best, the glaciers are broad and thin, and flow only by virtue of the steep gradients on which they lie. Some few, indeed, seem not to occupy valleys at all but to be virtual ice caps in form. The névé line is at about 2300 feet.

NORTH AMERICA

Arctic Islands and Labrador

Among the many large Arctic islands of North America only the eastern islands (Ellesmere, Axel Heiberg, Devon, Bylot, Baffin, and possibly Meighen) have glaciers today. The eastern islands reach high altitudes whereas the central and western islands are comparatively low. Thus there is a direct correlation between the presence of highlands and

the existence of present-day glaciers. The combined area of all these glaciers has been estimated to be 40,000 square miles.

Ellesmere Island (lat. 76° to 83°) the most northeasterly of the islands, is mountainous from end to end, with altitudes that are believed to reach 13,000 feet. It has a number of plateau ice caps of unknown extent, with outlet glaciers, some of which reach tidewater, as well as independent valley glaciers. Though very little is known about any of these ice masses, they appear to be thin, as is the northern part of the Greenland Ice Sheet immediately to the east. Furthermore, the northern part of the island (known as Grant Land) has the least ice. This fact appears to reflect an unfavorable position with respect to source of moisture.

At least the higher parts of Axel Heiberg Island, reaching over 6000 feet, carry small glaciers. Whether Meighen Island has an ice cap is uncertain.

The eastern part of Devon Island, 4000 to 5000 feet in altitude, appears to be covered with a large plateau ice cap. Many outlet glaciers flow outward from it, reaching the sea at a number of points.

Bylot Island (maximum altitude 6000 ft.) has at least one plateau ice cap. Baffin Island, the largest of the group, is high throughout its eastern part, with summits reaching at least 10,000 feet. Glaciers, mainly in the form of plateau ice caps lie near the southeastern tip of the island at an altitude of only about 3000 feet. Glaciers are larger and more abundant in the northern than in the southern part of the island.

In northernmost Labrador, only a short distance south of the southeastern tip of Baffin Island, the Torngat Mountains harbor a few very small glaciers of the cirque type.[42] Apparently they persist only because they occupy positions especially well protected from ablation. These glaciers are the only ones in eastern continental North America.

Alaska and Continental Canada

Pacific Coastal Mountains

The coast ranges of the Alaska-Yukon region are the site of most of the glacier ice and all the large glaciers of continental North America. Glacier ice is most extensive along the Gulf of Alaska coast, where both snowfall and average height of land are greater than elsewhere in the Alaska-Yukon region. The cool summers of the coastal area are a large factor in reducing ablation and thus help to maintain abundant glacier ice. The ice is an almost continuous network consisting of three kinds of glaciers. (1) Valley glaciers are the most numerous. Some of them

[42] Coleman 1921.

are very large, the largest known being Hubbard Glacier, 75 miles in length. (2) Broad glaciers, transitional between valley glaciers and ice caps, fill upland and intermont basins. The most conspicuous is the glacier occupying the broad depression between Mt. St. Elias and Mt. Logan. (3) Along the coast itself are piedmont glaciers, of which the largest are the famous Malaspina and Bering glaciers.

This network extends through several more or less distinct ranges immediately along the coast — the St. Elias Range (maximum altitude nearly 20,000 ft.), the Chugach Range (13,000 ft.), and the Kenai Range (more than 5000 ft.). Back of these the network continues through another tier of mountains — the Wrangell Mountains (maximum 16,000 ft.), the Alaska Range (20,000 ft.), and the Aleutian Range (10,000 ft.). As the Aleutian Range is followed southwestward along the Alaska Peninsula the glaciers become much less continuous and are confined to small valley glaciers on the higher peaks. The higher volcanic peaks of Unimak and Unalaska islands, the two most easterly islands of the Aleutian Chain, also carry small glaciers. Rise of the regional snowline away from the coast is shown by the fact that on the outer coastal flank of the St. Elias Mountains it is at 2300 feet but rises to 6500 feet on the eastern flank of the same mountains.

South of latitude 60°, the boundary between Yukon Territory and British Columbia, the glaciers of the coastal mountains become rapidly smaller and fewer. In the region between the Stikine River and the Skeena River the mountains are lower and the glaciers are correspondingly restricted. In southwestern British Columbia, however, inland from Vancouver Island, high individual mountain masses, such as that of Mt. Waddington, are the loci of numerous glaciers.

Interior Mountains

Away from the Pacific coastal region, even on high mountains, glaciers are few and small. The group of ranges together known as the Brooks Range in northeastern Alaska, with altitudes exceeding 8000 feet, carries small glaciers. Farther south a few glaciers occur in the high part of the Mackenzie Mountains along the Yukon-Northwest Territories boundary, at latitude 63° and reaching altitudes of more than 9000 feet. The Rocky Mountains Front Range between 51° and 54° has small glaciers, as has the Selkirk Range north of the United States–Canadian boundary. This poor development of glaciers in the interior reflects the reduced snowfall, and the increased ablation resulting from clearer summer skies, of this region as compared with the coastal region. Somewhat lower altitude in the interior may be a factor also.

United States

Mountains of the Pacific Coastal Region

Increased temperatures and decreased precipitation with decreasing latitude are the principal causes of the small development of glaciers in the United States as compared with Canada. The Olympic Mountains west of Puget Sound and the high volcanic peaks of the Cascade Mountains in Washington and Oregon (ending southward with Mt. Shasta in northern California) carry small valley glaciers, most of which are confined to the valley heads. The greatest single development of glaciers in this region is the cluster on Mt. Rainier. Farther south, on the Sierra Nevada, separated from the coast proper by lower mountains, there are about 50 small glaciers occupying protected positions in the Mt. Whitney district.

Rocky Mountains

Farther in the interior, glaciers are nearly negligible. They are so small that they are confined almost entirely to cirques opening to the north and east. Such small ice masses occur in the Lewis Range in the Glacier Park district of Montana, and the Wind River Range and Teton Range in western Wyoming. Farther south other glaciers are found in the Medicine Bow Range in southern Wyoming, and in the Front, Sawatch, and Sangre de Cristo ranges in Colorado and New Mexico. What are possibly the southernmost glaciers in the United States occur in the Sangre de Cristo Range at latitude 37°35′.

CENTRAL AND SOUTH AMERICA

Mexico

Between the United States and the north coast of South America there are only two peaks high enough to carry extensive snowfields in the prevailingly dry climate of the Mexican Plateau. These are the volcanic cones Popocatepetl (about 18,000 ft.) and Ixtaccihuatl (about 17,000 ft.), standing close together in the region southeast of Mexico City. Ixtaccihuatl has a few small glaciers.

South America

Existing glaciers in South America are confined to the Andes Cordillera, but they extend from near the Caribbean to near Cape Horn. Their average altitudes descend from more than 18,000 feet near the

equator to 3000 feet in Tierra del Fuego. In southern Chile the glaciers include extensive plateau and intermont ice caps. Farther north only valley glaciers are present, including broad thin glaciers that lie on the slopes of volcanic cones and that strictly speaking do not occupy valleys at all. The combined area of existing glacier ice in South America is estimated at 9600 square miles.

In *Venezuela* glaciers occur on the Sierra Nevada de Mérida (lat. 9° N) at about 15,000 feet.

In *Colombia* groups of glaciers occur on several high mountains. The farthest north, directly on the Caribbean, is the Sierra Nevada de Santa Marta (lat. 11° N) where the regional snowline lies at 15,000 to 16,000 feet. Others include the Sierra Nevada de Cocuy (6° N), Nevado del Tolima (5° N), Nevada del Huila (3° N), Sierra Nevada de los Coconucos (2° N), Páramo de Almorzadero, Ruiz, and Páramo de Sumapaz (4° S). Summit altitudes generally range between 15,000 and 19,000 feet.

In *Ecuador* many high volcanic cones between Quito (0°) and Palmira (2° S) carry glaciers.

In *Peru* many of the highest peaks in the Cordillera Blanca and the Cordillera Huayhuash (8° S–12° S) are the sites of glaciers. Farther south, glaciers occupy the highest parts of the Cordillera de la Costa and, farther inland, the chain of mountains north and east of Lake Titicaca.

In *Bolivia,* also, isolated high peaks in the coastal cordillera and in the Cordillera Real, near La Paz, carry glaciers.

In *Chile* scattered high peaks harbor glaciers from latitude 18° S to about 34° S (the vicinity of Santiago). From here southward to about 41° altitudes are lower and there are almost no glaciers. South of latitude 41° glaciers become gradually larger and more abundant; between 46° and 52° they form a nearly continuous complex like the glacier complexes of coastal Alaska. The most southerly glaciers are on Hoste Island (Tierra del Fuego) at about 55° S.

EURASIA

Arctic Islands

Spitsbergen[43]

Spitsbergen (Norwegian *Svalbard*) is a group of islands in the Arctic Sea of which the four largest are West Spitsbergen, North East Land, Edge Island, and Barents Island. Their combined area is about 25,000 square miles, of which more than 22,000 square miles is estimated to be covered with glacier ice. The snowfall on Spitsbergen is derived chiefly

[43] See Sandford 1929*b*; G. W. Tyrrell 1927; Ahlmann 1933*b*.

from moist air in masses moving northeastward from over the Atlantic Ocean. Hence precipitation on the islands diminishes from southwest to northeast. However, temperatures diminish in the same direction to such an extent that glacier ice covers more of eastern than of western Spitsbergen.

West Spitsbergen is a much-dissected plateau with an average altitude of perhaps 2000 feet but with mountains that reach extreme heights of 5000 feet. The northeastern part of this island (the highest and coldest part) has an ice covering that is continuous. Conditions here approach those of an ice sheet. Elsewhere on the island the glaciers consist dominantly of valley glaciers and secondarily of small plateau ice caps, although in many districts these coalesce extensively. In the northern part of the island, outlet and valley glaciers reach the sea and in places form extensive piedmont glaciers. In the southernmost part valley glaciers terminate well above sealevel. The whole glacier system of West Spitsbergen is much like that in the coastal mountains of Alaska, though in Spitsbergen the glaciers are more closely spaced and cover a greater proportion of the land area. In northwestern Spitsbergen the altitude of the névé line varies between 1300 and 2000 feet.

Barents and Edge islands have glaciers only in their eastern parts, which are higher and colder than their western slopes.

North East Land is almost completely blanketed by two large ice caps and two or three small subsidiary ones, each of them dome-shaped, with extreme altitudes of 2000 to 2400 feet. The ice appears to be very thin, perhaps no more than 350 feet thick, and is very sluggish, flowing slowly in some sectors and in others being nearly or quite motionless. Precipitation is slight because West Spitsbergen lies between North East Land and the main source of moisture. The extreme cold reduces ablation to a small figure. In consequence the ice caps of North East Land, like the ice sheet of Antarctica, are being starved and are spreading out, thinning, and approaching a condition of stagnation. In comparison with the relatively high névé line of West Spitsbergen, the névé line of North East Land is low, ranging from 1700 feet in the southwest to 300 feet in the east.

Franz Josef Land

Franz Josef Land is an archipelago of more than 50 islands with a combined land area of nearly 7000 square miles, nearly all of which is covered with glacier ice. With few exceptions each island is a plateau, large or small, covered with its own ice cap. Maximum altitudes range from 2500 feet on the southeastern islands down to 1000 feet or less on

most of the others. The greatest single ice-free area, 60 miles long, is on the westernmost island, and probably exists because of exceptionally low altitude. The ice caps are thin, probably no more than 500 to 600 feet in their thickest parts, for they are seen to conform to the underlying topography and nunataks project through them in places. The rate of flow of the ice is inferred to be very slight.

Novaya Zemlya

About one-sixth (5800 square miles) of the area of Novaya Zemlya[44] is covered with glacier ice. Nearly all of it is on the northern half of the island, or rather double island, for the winding Matochkin Strait, barely 1500 feet wide, divides it into two parts.

Near this strait, valley glaciers flow out of mountains 3000 feet high, but they do not reach sealevel. The névé line here lies at 1800 to 2000 feet.[45] Progressively toward the north the glaciers increase in number and length. At about latitude 75° they merge into a continuous ice cap covering the plateaulike surface of the island. This cap exceeds 3000 feet in greatest altitude and has many outlet glaciers that reach the sea on wide fronts. The névé line descends, correspondingly, to 1300 feet at Cape Mauritius.[46] The ice is generally thin. Its thickness averages less than 800 feet and reaches an estimated maximum of no more than 1400 feet.[47]

Severnaya Zemlya

The Severnaya Zemlya archipelago consists of three large islands and some much smaller ones, all with a combined area of about 13000 square miles. Their terrain consists of dissected plateaus with maximum altitudes about 1500 feet and average altitudes of less than half this figure. About 42 per cent of the land area is covered with glacier ice[48] largely in the form of small ice caps. The islands lie along a northeast-southwest axis. Komsomolets Island, on the northwest, is 65 per cent ice covered. October Revolution Island, in the middle, is 45 per cent covered, while Bol'shevik Island, on the southeast, is only 21 per cent covered. All three islands have much the same range in altitude. This difference seems to be the result of their relative position with respect to source of snowfall. Moisture is brought to the islands chiefly from the west.

[44] Grönlie 1930.
[45] Nordenskjöld and Mecking 1928, p. 157.
[46] Hess 1904, p. 111.
[47] Gerasimov and Markov 1939, p. 56.
[48] Gerasimov and Markov 1939, p. 56.

The ice caps are said to be no more than 600 to 800 feet thick, and the névé line is not far above sealevel.

To the northeast, Bennett Island (lat. 76°40′ N, long. 149°00′ E) carries an ice cap.

Continental Eurasia

Scandinavian Mountains

The Scandinavian Mountains form a continuous chain 1000 miles long, extending from the Barents Sea to the North Sea and throughout much of their length forming the boundary between Norway and Sweden. Summit altitudes in their northern part range between 3000 feet and 6000 feet, and increase southward, reaching an extreme of 8000 feet in southern Norway. From their northern end, near latitude 71°, southward to about 60° these mountains are marked by scattered glaciers. Some are small valley glaciers, but many are small ice sheets that cap plateau areas between valleys and that drain downward through small outlet glaciers. The largest (Jostedals Glacier between lats. 61° and 62°) is 35 miles long. In general the glaciers are most numerous and largest where mountain altitudes are highest. Their altitudes (and with them the altitudes of the névé line) range from 3000 feet in the northern coastal region to as much as 7800 feet in the interior of southern Norway. Their combined area is estimated at 2416 square miles. Most of the glaciers lie in Norway, but some are on the Swedish side of the mountain watershed.

The Alps[49]

The region of the Alps is dotted with glaciers from French Savoie on the west to the region south of Vienna on the east, an overall distance of more than 350 miles. The glaciers (including very small masses that may be only thin accumulations of névé) number more than 1200 and have a combined area estimated at more than 1900 square miles. The regional snowline varies between 8000 feet (generally in the west) and 9500 feet (generally in the east). The larger glaciers are confined to the western and central Alps (Mont Blanc Group, Pennine Alps, and Bernese Oberland). The largest individual glacier is the Aletsch, more than 17 miles in length. Many glaciers head not in cirques but in upland basins filled with névé.

[49] Penck and Bruckner 1909.

Pyrenees

Glaciers occur on the northern slopes of the central Pyrenees, the highest part of the chain where summit altitudes locally exceed 11,000 feet. The glaciers are of the valley type and are small, having a combined area of only 15 square miles. The regional snowline varies between 8800 and 9200 feet.

Western and Central Asia

In western Asia the higher parts of the Caucasus have a glacier complex resembling that of the Alps. There are recorded more than 800 glaciers, most of which are very small.[50] Some of the high ranges south of the Caucasus, and the Elburz Mountains along the southern coast of the Caspian Sea, also have valley glaciers.

In central Asia the Pamir Mountains in the Tadzhik Soviet Republic and in the western border of Sinkiang harbor nearly 1200 glaciers, exclusive of cirque glaciers, and including some of the largest valley glaciers in the world.[51] Despite the interior position of these mountains the abundant snowfall brought by monsoon winds from the Indian Ocean maintains a vigorous glacier regimen, with upland intermont glaciers and snowfields as well as valley glaciers. The regional snowline is in the neighborhood of 18,000 feet.

Monsoon-wind nourishment accounts also for the extensive present-day glaciers in the high eastern part of the Hindu Kush, south of the Pamirs on the border between Afghanistan and India, as well as for those of the Himalaya Range (snowline somewhat more than 16,000 ft.) and the Karakoram and other great ranges farther north in the Himalayan arc in southern and eastern Tibet (snowline 18,000 to 20,000 ft.).

Glaciers occur in the K'un Lun-Nan Shan chain in northern Tibet and southern Sinkiang, and in the high mountains of southeastern Tibet, Sikang, and Yunnan. Farther north the T'ien Shan chain in northwestern Sinkiang and the adjacent part of the Kazakh Soviet Republic carries glaciers. To the northeast small glaciers occur in the Altai-Sayan mountain system in northwestern Outer Mongolia, the Tannu Tuva Republic, and Siberia. The glaciers in the Sayan Mountains are reported to occur at altitudes of 10,000 feet and higher.

Detailed information on the character and extent of most of the glaciers in central Asia is lacking.

50 Hess 1904, p. 87.
51 Korshenevsky 1930.

Northeastern Asia

As far as is now known, only four highlands in northeastern Asia carry glaciers, all small and of the valley-glacier type. The Chersky Mountains lying west of the Indigirka River have a few glaciers at and above 8500 feet in the sector just north of the Arctic Circle. Some of the high volcanic peaks on the Kamchatka Peninsula (the highest reaches nearly 16,000 feet) have small glaciers at and above 6000 feet. The Koryak Range along the Bering Sea coast north of latitude 60° has small glaciers at 3000 to 4000 feet. The Anadyr Mountains in extreme northeastern Siberia have a few glaciers in protected positions at 3000 to 4000 feet.

NEW GUINEA

Central New Guinea is the site of a high mountain range dominated by the Carstensz Group, of which the highest peak, Ngga Poeloe, (lat. 4°05′ S, long. 137°10′ E) reaches an altitude of 16,600 feet. This group harbors a plateau ice cap at least 5 miles in longest diameter with one well-developed outlet glacier, and another ice cap about 2 miles long with three outlet glaciers.[52] The regional snowline lies near 16,000 feet. Idenburg Peak, 10 miles west of the Carstensz Group, and Mt. Wilhelmina (lat. 4°15′ S; long. 137°10′ E; alt. 15,675 ft.), the high point of the Emma Mountains, are also the sites of small glaciers. All these glaciers are thin, but, despite the nearness of this district to the equator, they are maintained because of abundant precipitation from the moist maritime atmosphere.

NEW ZEALAND

The North Island of New Zealand has only one peak high enough to maintain glaciers. This is the volcanic cone Ruapehu (lat. 39°20′ S; alt. 9175 ft.) in the Tongariro National Park, whose mile-wide summit crater contains a glacier.

On the South Island, however, the high and nearly continuous Southern Alps, dominated by Mt. Cook (12,349 ft.) are marked by a total of more than 50 valley and intermont glaciers and small plateau ice caps through a distance of more than 200 miles (Fig. 67). The longest valley glacier, the Tasman, is 18 miles in length. The larger and more favorably situated glaciers reach down to within 400 feet of sea-level. The combined area of glaciers on the South Island is believed to be nearly 400 square miles, and summits less than 7000 feet high in the southern part of the highland carry glaciers.

[52] Dozy 1938.

The ice is maintained by abundant snowfall precipitated from moist westerly winds. The regional snowline rises from about 6000 feet on the west flank of the mountains to 7500 feet or more on the east flank.[53]

SUBANTARCTIC ISLANDS

Most of the subantarctic islands have glaciers at the present time, and some of them are very extensively ice covered. Details are given in Table 17.

EAST AFRICA

East Africa has three peaks which, though situated close to the equator, are high enough to rise above the regional snowline. These are:

Peak	Latitude	Longitude	Altitude (feet)	Regional Snowline (feet)
Ruwenzori	0°24′ N	29°54′ E	16,912	15,180
Kenya	0°10′ S	37°18′ E	17,143	16,335?
Kilimanjaro	3°05′ S	37°22′ E	18,045	17,820

All three peaks carry glaciers today. There are ten on Mt. Kenya alone.[54]

SIGNIFICANCE OF DISTRIBUTION

The foregoing brief summary of the distribution of existing glaciers indicates that, as the regional snowline descends with increasing latitude, so do glaciers exist at low altitude only in the Arctic and Antarctic regions. In extremely high latitudes, under conditions of intense cold, glacier regimens are inactive: nourishment, ablation, and rates of flow are small. Such are the conditions encountered in the Antarctic Ice Sheet, in the northern part of the Greenland Ice Sheet, and on some of the islands of the Arctic Sea.

In middle and low latitudes, on the other hand, glaciers occur only relatively far above sealevel. How far above depends largely on their position with respect to sources of moisture. In regions with a maritime climate glaciers are maintained at much lower altitudes than in regions with a continental climate, because snowfall is abundant and because ablation is greatly reduced by the cloudiness that accompanies maritime climates. If mountains are high enough, as in East Africa, South America, and New Guinea, glaciers can be maintained near and even directly on the equator.

[53] Hess 1904, p. 102.
[54] Nilsson 1940, p. 63.

Chapter 5

GLACIAL EROSION AND TRANSPORT

INTRODUCTION

As we have seen, the evidence that led Venetz, Charpentier, and Agassiz to realize that glaciers in the Alps had formerly spread far and wide was the presence, in areas far removed from existing glaciers, of features that these men recognized as being indubitably the work of glaciers. These features were of two kinds: first, boulders and other deposits consisting of rocks foreign to the places where they occur; and, second, bosses of grooved, scratched, and polished bedrock. Although many other evidences of glaciation have been recognized since, by and large these two basic features are still the cornerstones of our mapping of the glaciated areas of the world, and as such they deserve to be considered in some detail. Therefore most of this chapter and the one following it are devoted to the markings and the land forms made by glacial erosion. Glacial deposits are discussed in Chapters 7 and 8.

By *glaciation* is usually meant the alteration of any part of the Earth's surface (usually by means of erosion or deposition) in consequence of glacier ice passing over it.[1] Erosional alteration may consist of anything from the inscribing of minor scratches to the profound excavation of valleys; depositional alteration may range from the lodgment of a single foreign stone to the building of a thick and extensive mantle of glacial sediments. By *deglaciation* is meant the uncovering of any area as a result of glacier shrinkage.

SMALL-SCALE FEATURES OF GLACIAL ABRASION

Over wide areas in glaciated regions the bedrock has been markedly abraded by the basal part of the ice shod with rock fragments of all sizes, and in the process the rock fragments themselves have become

[1] C. S. Wright and Priestley (1922, p. 134) distinguished between *glaciation* ("the erosive action exercised by Land-Ice upon the land over which it flows") and *"glacierization"* ("the inundation of land by ice . . ."). The usefulness of this distinction is questionable. *Glacier covered* is a clearer and simpler term than *glacierized* for describing areas now covered with glacier ice. As for areas formerly so covered, they must be glaciated in the sense used in the text above; otherwise the fact of former overspreading by ice could not be established. Therefore a term such as *glacierization* seems unnecessary.

abraded. The bedrock has been partly blanketed by glacial and other deposits which conceal the ice-abraded surfaces, but usually enough are exposed to permit determination of the extent of glaciation. On the other hand mechanical wear of rock on rock is caused by several quite different natural processes, and some of the resulting abraded surfaces are so much alike that the process responsible is not always identifiable. Supporting evidence of other kinds that glaciation has occurred has to be sought before the fact of glaciation can be established.

Striations, Polished Surfaces, and Grooves

The most common and conspicuous unit of glacial abrasion is the striation (stria or scratch). Chamberlin[2] described striations succinctly as "fine-cut lines on the surface of the bedrock which were inscribed by the overriding ice." The fineness of the scratch reflects the size of the abrading fragment. The finest scratches were probably cut by grains of silt and sand, not imbedded in the basal ice because the pressure exerted would have retracted them into the ice, but caught between the bedrock floor and large pieces such as cobbles and boulders in the base of the glacier. These delicate striations usually appear only on fine-grained and mineralogically soft rocks like limestones and shales. On harder and coarser-grained rocks such as granites only the coarser striations, probably made by pebbles, are found.

The finest striations grade downward into a general polish, whose smoothness and brilliance are limited only by the fineness of the abrading grains of silt and by the textural and mineralogic ability of the abraded rock to take a polish. Most polished surfaces also bear readily visible scratches, because it is rare indeed that all the fragments in the base of the glacier lie within a single narrow range of grain size.

In the other direction striations grade upward into grooves—giant furrows that have been observed only in soft rocks such as limestones and shales. The grooves appear to have been made by the fortuitous enlargement of single striations as angular pieces of rock carried in the base of the ice gouged them deeper. The basal ice, under heavy pressure, molded itself to fit the slight depressions thus made and further localized the abrasive process. As the long, straight depressions enlarged, the plastic ice flowed into them, fitting them tightly. Individual grooves reach depths of several feet and lengths of hundreds of feet. Many have overhanging sidewalls, and on the under sides of some of the overhanging walls are delicate striations, fully parallel, of course, with the groove. A famous locality for grooves is Kelleys Island, Ohio, where a

[2] T. C. Chamberlin 1888, p. 216.

broad horizontal surface of limestone subjected to vigorous glaciation furnished ideal conditions for striation and grooving.[3]

Relation of Striations to Glacier Flow

Not all striations on rock are of glacial origin; agencies other than glacier ice can and do make striations. A common nonglacial agent in high latitudes is floating ice in rivers, lakes, and the sea.[3a] This fact was recognized in both Europe and America even before the glacial theory was thought of. At that time striations were attributed to the action of floating Arctic ice during a universal submergence of the lands. Except in ideal instances no certain means of distinguishing between glacier-ice scratches and floating-ice scratches on the rocks has been discovered. This uncertainty leaves us in serious doubt as to the directions of flow of former glacier ice in Arctic Canada, much of which was deeply submerged during the close of the last glaciation, because many of the striations within the submerged region may have been made by floating ice. Hence the broad problem of glacier-ice movement in that region is still unsettled.

Striations have been made also by snowslides, by landslides, and, most remarkable of all, by *nuées ardentes* — dense clouds of white-hot solid particles that roll down the slopes of a volcano during certain intensely explosive eruptions.

Thus it appears that under suitable conditions striations can be made by any heavy mass flowing according to the laws of fluid dynamics. This means that reasonable evidence of glacial origin must be taken before a striation is interpreted as glacial. It means also that a rigid grip on the abrading tool by the flowing medium is not necessary in order to produce a scratch on the rock beneath. However, a single stone in the sole of a glacier is under such great pressure that it is soon retracted (by pressure-melting and refreezing) into the body of the ice so that it ceases to touch the rock floor. If the glacier were shod with only a few stones, such retraction would prevent the engraving of any long striations. Long, continuous scratches are probably made by stones in a very abundant basal load. The high proportion of stones to ice makes the whole basal mass so rigid that retraction can not occur and continuous striation results.

In considering a striation that is actually of glacial origin, we must determine its trend or compass bearing, but we must also look for evidence as to which way the abrading ice flowed. Evidently there are

[3] Carney 1909*a*.
[3a] J. W. Dawson 1893, pp. 105–110.

two possibilities. Was the glacier that made a north-south striation flowing northward or southward? In the vast majority of instances the actual direction of flow is not determined, or is not even determinable, from evidence furnished by the striation itself. It is inferred from other features in the vicinity. A striation in itself ordinarily does not indicate direction of flow; it indicates two possible directions. Although some striations are unsymmetric, being blunt at one end and tapering at the other, this fact does not indicate direction of flow. Unsymmetric striations on stoss slopes[4] usually lie with their blunt ends in the downstream direction, apparently as a result of gouging, whereas those on lee slopes usually are blunt at the upstream end, apparently as a result of quarrying.[5] In general, striations are much more abundant on stoss surfaces than on lee surfaces. We are obliged to conclude that the majority of striations are not as specifically useful direction indicators as has been thought. Undoubtedly some north-south striations thought of as made by south-flowing ice because other features in their vicinity indicate such flow actually were cut by ice flowing northward during a late phase in the glacial history. This matter is discussed in a later chapter.

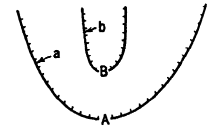


FIG. 12. Radial "lines of flow," *a*, in a glacier lobe (*A*) in a vigorous condition, and (*b*) in a glacier lobe (*B*) in a shrunken condition. Sets of striations formed parallel with these "lines of flow" would have different trends. (After T. C. Chamberlin.)

At first thought it might seem that all the striations within a single region would be likely to show parallelism, but this is not true. Even though the rock surface beneath the flowing ice were perfectly flat the scratches made on it near the margin of a glacial lobe would be ideally radial (Fig. 12), and the true picture of the movement of the glacier could not be reconstructed until the scratches within the whole area of the former lobe had been mapped. They would diverge from each other. Even greater divergence in the trends of the striations can be caused by topographic irregularities. That glacier flow can be deflected by very minor irregularities has been demonstrated in Glacier National Park, Montana, where striations show that eddylike flow was imparted to former glacier ice by minor obstacles.[6] Wherever an ice sheet flowed across much-dissected terrain the striations show that the ice in the main valleys tended to follow the valleys, even though the trends of these depressions depart from the general direction of ice flow by a wide angle.[7] The glacial scratches in the New England region (Fig. 13) show

[4] Slopes facing the oncoming ice.
[5] H. C. Lewis 1885, p. 557.
[6] Demorest 1938; see also Holmes 1937.
[7] T. C. Chamberlin 1888, p. 197.

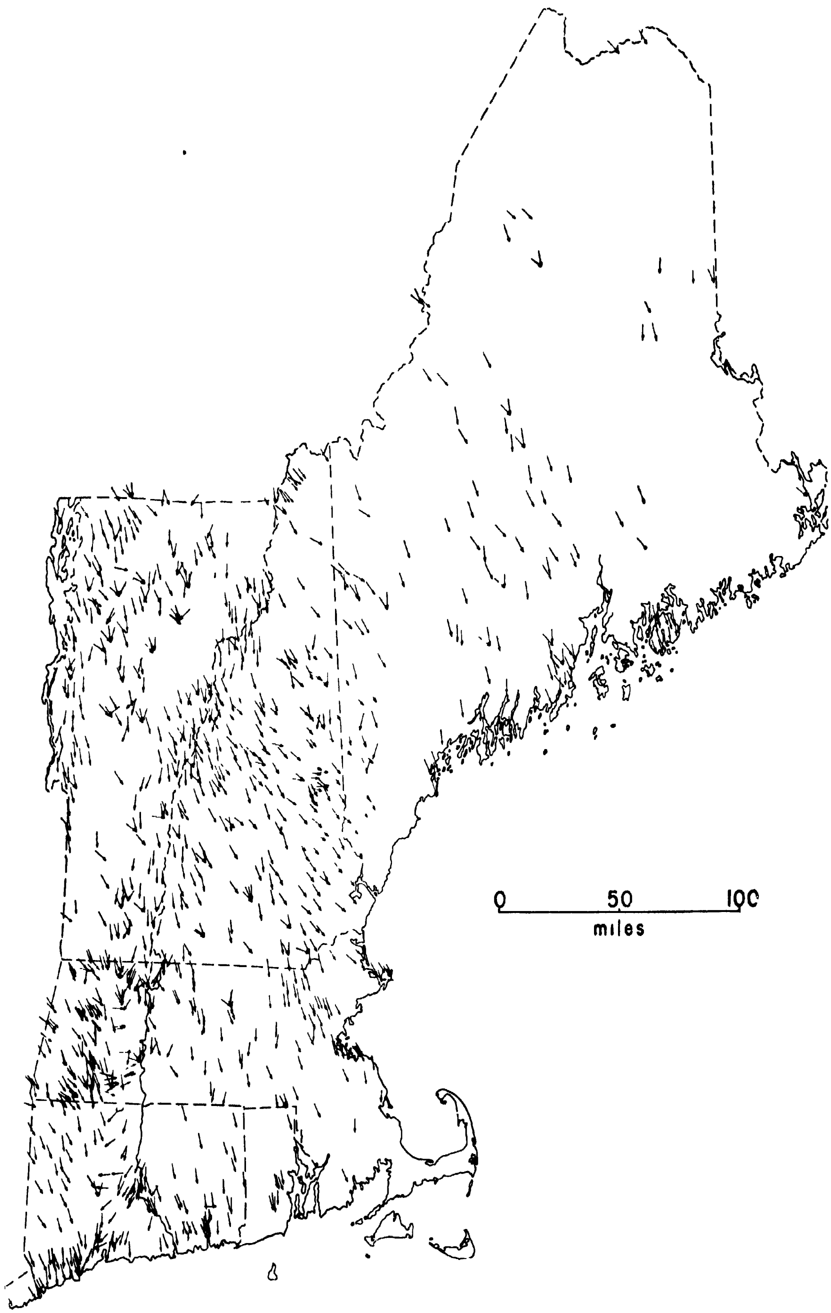


FIG. 13. Glacial striations in New England, compiled from many sources.
(J. W. Goldthwait.)

this relationship distinctly. In central Connecticut striations on the higher hills indicate a regional trend of glacial flow that is somewhat south of east. Striations in the valleys, however, parallel the valleys and diverge from the regional trend by as much as 90°. Similarly in the Presidential Range in New Hampshire[8] high-level striations record a regional trend of flow of about S 40° E, but striations in the valleys diverge by as much as 50°.

Summarizing the foregoing discussion, we conclude that (1) not all natural striations on rock are of glacial origin, (2) few striations indicate in which of two directions the ice flowed, and (3) strong divergences in the trends of striations are produced by lobation of glacier ice and by topographic irregularities. All this adds up to the fact that striations must be interpreted with a great deal of selective judgment. A few striae mean little, but the "weight of evidence" of many hundreds of striae, accurately mapped, ordinarily yields a fair picture of the movement of the ice concerned, as shown by Fig. 13.

Crossing striae — two or even three sets of striations with discordant trends — are not uncommon, and it is usually possible to determine the relative ages of the different sets. Many of these have been interpreted as evidence of two or more distinct glacial ages, and for some this interpretation is undoubtedly justified. However, crossing striae can be caused by change in direction of flow of a single lobe of an ice sheet (Fig. 12) as a result of (1) shrinkage involving backward shifting of the ice margin, (2) thinning, causing topographic irregularities to exert increased influence, and (3) shift in the position of the center of outflow of ice. Therefore evidence of a long lapse of time between the making of the successive sets of striae is prerequisite to the interpretation of crossing striae as the product of successive glacial ages.

When plotted on a continental scale, striations are seen to be very irregularly distributed. The spotty distribution, well shown on the Glacial Map of North America,[9] is the result of a number of highly variable factors such as (1) lithology, (2) destruction by postglacial erosion, chiefly weathering, (3) concealment beneath a blanket of drift, (4) debris content of the glacier ice, (5) variable observation and mapping. It is difficult to say which of these factors is the most important, although in North America 5 would probably rank first. Countless numbers of striae exposed to view on the resistant rocks of the Canadian Shield have never been recorded or even observed by geologists. Probably 3 would rank second in importance.

Marked local variations in distribution of striations are on record. At

[8] R. P. Goldthwait 1941.

[9] Flint and others 1945.

Kelleys Island, Ohio, stripping operations removed the till overlying the limestone bedrock and exposed a smoothly polished area more than 800 feet long lying between two rough areas bearing no trace of glacial erosion. Beyond one of the rough areas is a larger surface showing deep grooves and striations. All these areas lie in about the same plane. The basal ice is held to have become so heavily loaded with rock fragments that it stagnated in places, the ice above flowing over it and eroding the bedrock beyond.[10]

Friction Cracks[11]

Commonly associated with the striations on hard brittle rocks are small transverse markings, some of them crescentic, which have been variously termed crescentic fractures, crescentic gouges, and chattermarks. Because of confusion as to the origin of these markings and as to their significance in determining the direction of flow of the ice responsible for them, the term *friction crack*[12] has been used.

A friction crack is a glacially made crescentic marking having a distinct fracture that dips forward into the rock. It is so named because all such cracks are believed to be made by local increase in friction between ice and rock. Some are cracks only; others have depressions showing that rock has been chipped away along the line of intersection of the crack with the bedrock surface.

Friction cracks are 1 to several inches in length and occur on comparatively flat-lying rock surfaces, more commonly on the stoss sides than on the lee sides of hills. They are likely to occur in sets, the fractures constituting each set lying in series in the direction of ice flow. Some sets are traceable through distances of several yards. They are best developed in hard, brittle rocks, especially quartzite (Fig. 14).

Mechanically, friction cracks are produced in at least two different ways, which are discussed in detail by Gilbert[13] and Harris.[14] Some cracks are convex downstream; others are convex upstream. Hence orientation of the convexity of the crack is not a sure indicator of the direction of glacier flow. However, it appears probable that, regardless of surface orientation, all friction cracks dip forward (downstream) into the rock. If this is so, then the dip of a crack is a useful indicator of direction of flow. The dip of a friction crack has the further advantage that it can be recognized and measured even after the bedrock has been weathered sufficiently to destroy all striations.

10 Carney 1909*a*, p. 640.
11 Gilbert 1906; Harris 1943 and references therein.
12 Harris 1943.
13 Gilbert 1906.
14 Harris 1943.

Chattermarks are not friction cracks, for they possess no fracture. They are merely the scars made in series by vibratory glacial chipping. They were named in analogy with a machinist's chattermark, which results when a tool, not firmly held, plows across a piece of metal. Each

J. D. Bateman, Geol. Survey of Canada

FIG. 14. Friction cracks in quartzite, Granville Lake district, Manitoba. The ice that made them flowed away from the observer.

scar marks the place where rapidly accumulated stress in the metal was relieved by failure of the metal. Glacial chattermarks possess no consistent form and are not in themselves reliable indicators of direction of ice movement.

QUARRIED SURFACES

Many slopes facing in the downstream direction and underlain by rock cut by joints, sheeting, and stratification planes are roughened by

glacier ice to profiles that are irregular and even sharply steplike (Fig. 17). They are confined to bosses and small hills and are not developed on slopes of mountainous dimensions. Monoliths up to many feet in diameter, bounded by structural surfaces, are lifted from the rock by the flowing ice, carried forward (rather than upward), and removed. The process has been termed plucking, and also quarrying, in analogy with the lifting out of blocks in a stone quarry. It is generally believed that the force of flow of the ice is adequate to this operation, although it has been held that pressure-controlled freezing of melt-water in the rock immediately beneath the glacier is an important factor.[15]

Not only does quarrying contribute largely to the stones present in the till, but also the quarried slopes are a helpful indication of the direction of flow of the ice. They indicate which of the two directions recorded by the striations was taken by the ice. The persistently asymmetric arrangement of bosses and small hills in a strongly glaciated district, each hill having a comparatively gentle abraded slope on the stoss side and a somewhat steeper and rougher quarried slope on the lee side, is termed *stoss-and-lee topography.*

TRANSPORT

Data from Existing Glaciers

A common fact of observation is that rock fragments in transport by glaciers are concentrated predominantly close to the contact between glacier and bedrock. This is clearly seen where the basal parts of many glaciers are exposed at their termini. It is even more evident along the lateral margins of valley glaciers where ablation has exposed continuous belts of debris-filled ice ("dirty ice") including rock fragments acquired by abrasion and quarrying, augmented by other debris fallen from above. These belts are *lateral moraines.* They have a ridgelike surface form, but in a great many lateral moraines the ridge consists of a surface veneer of debris covering a ridge of "dirty ice." The belt of debris has protected the ice beneath it from ablation, thus forming a ridge that stands above the more rapidly wasting clean ice near by. As the ridge steepens, the debris slides away, exposing the ice to wastage. Thus the lateral moraine gradually widens and hides from view the lateral contact of ice with bedrock.

Away from this contact rock debris is usually so scanty that the ice, with certain exceptions, seems almost pure. One exception consists of

[15] Holmes 1944*a*.

scattered debris dropped, rolled, or slid out onto the surface of a valley glacier from inclosing cliffs, and fine-grained debris blown by the wind and lodged on the surface in the terminal zone. Another exception consists of debris formerly imbedded in the ice but brought to the surface in the terminal zone by gradual ablation. In some glaciers such material in time forms an almost complete coating over the terminal zone and greatly retards further ablation. A third and most conspicuous exception consists of medial moraines.

Medial moraines parallel the lateral moraines but lie nearer to or at the center line of the valley glacier. As they are merely the lateral moraines of tributary glaciers extended down the main, their number depends on the number of tributaries of which the main glacier is built. Figure 15 shows a main glacier whose medials are so numerous that

FIG. 15. Barnard Glacier, Alaska, showing relation of lateral and medial moraines to tributary glaciers. Kame terraces or ablation moraine or both are developing along margins of main glacier. The spurs flanking the main valley are faceted. The high peak is Mt. Natazhat (13,480 ft.), 25 miles away on the Alaska-Yukon boundary.

Bradford Wash

their exact number is difficult to determine. At least thirteen tributaries are visible, and presumably there are many more out of sight. It should be emphasized that medial moraines are rarely mere surface accumulations of rock waste. They are the surface outcrops of debris-rich zones in the ice which usually extend right down to the bedrock floor beneath

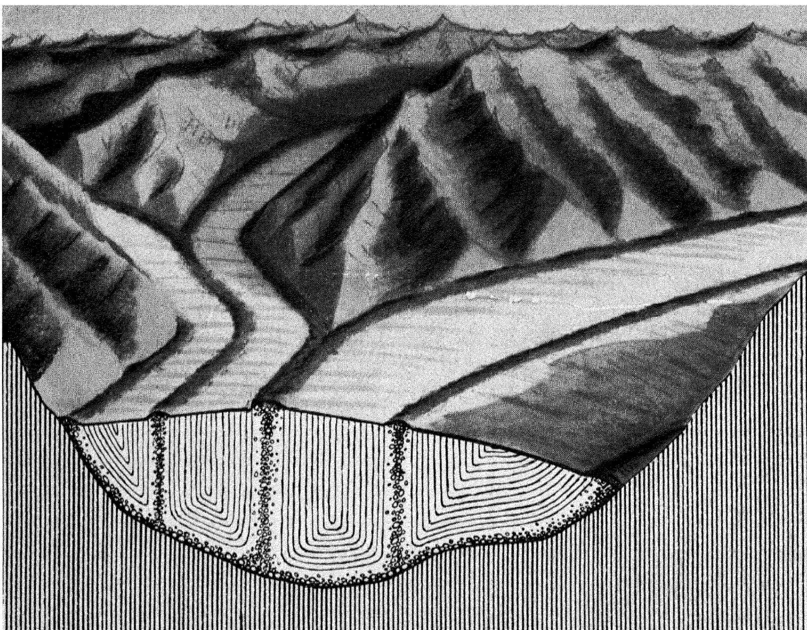


FIG. 16. Perspective view and cross section of ideal valley glacier, showing relation of surface moraines to tributary glaciers. (Modified after Martonne.)

the glacier. In other words the tributaries maintain their individuality as distinct parts of the main through long distances downstream from their junctions with the main. This is well brought out in a sketch (Fig. 16) adapted from Martonne. A glacier with several medial moraines therefore is likely to be transporting a very considerable volume of rock debris which, with increasing ablation of the ice toward the terminus, covers the surface in the terminal zone nearly completely. Here ridging of the ice through the protective action of surface debris becomes very complex, and the resultant lateral sliding of rock fragments aids their spreading over the surface. Areas of ice not coated with debris melt down to form basins which fill with meltwater and accumulate sediment, only to be soon destroyed. The individual medial moraines lose their

identity and merge into an irregular superglacial veneer of rock fragments, referred to as *ablation moraine.*

Linear surficial moraines develop on an ice sheet only where the ice is cut by nunataks or is subdivided into outlet glaciers. Probably quarrying takes place on the lee slopes and tops of steep hills that project upward into the ice sheet without being high enough to break the surface of the ice.

Data from the Drift

The significant inferences drawn from study of the drift are discussed in Chapters 7 and 8, but certain facts about the drift are so important to an understanding of glacial transport that they must be mentioned here.

One fact is that glaciers carry very little rock material far beyond the places where it is picked up. This is demonstrated by the almost universal close relation between the composition of the drift and that of the local bedrocks. It fits, also, the concept developed above, of a glacier loaded with drift only where it is in contact with the ground. The great bulk of the flowing ice has little opportunity to acquire a load of rock fragments. The load in the basal ice is ordinarily so abundant that the rate of attrition is high, and most of the component fragments literally wear out within a short distance of travel. Furthermore, within the terminal zone of the ice, opportunity for deposition is large, so that here much of the rock material becomes lodged on the ground very soon after it is picked up.

A very small proportion of the rock matter picked up does, nevertheless, travel long distances. Stones and boulders from Scandinavia and Finland were carried hundreds of miles by the Scandinavian Ice Sheet to eastern Britain, Germany, Poland, and (700 to 800 miles) to Russia. Boulders from Quebec were carried by the Laurentide Ice Sheet as far as 600 miles to positions in Kentucky, Indiana, Illinois, and Missouri. Most such stones and boulders consist of durable rock types, with a large proportion of hard resistant minerals, and with no joints or other planes of weakness in them. Apparently they survived long-distance travel in the base of the ice at the expense of considerable loss of size by attrition, though some of them may have traveled in englacial positions where there were few other rock fragments to abrade them.

A second fact about the drift is the conspicuous lift given to pieces of rock that survive long enough to be carried great distances. Wherever the drift is studied in detail and the direction of flow was against the slope of the land, lifting of rock fragments in transport is evident. In

eastern Ireland boulders and shells in the drift occur many hundreds of feet above their known places of origin, in some cases less than a hundred miles away. Granitic boulders from lower land farther north occur high in the Adirondack Mountains, and the drift on the Allegheny Plateau in central New York includes abundant stones derived from the lowland north of the Plateau. Many stones and boulders on the Presidential Range in New Hampshire have been lifted 2000 to 4000 feet above their highest possible places of origin in the bedrock. Drift deposited by the Laurentide Ice Sheet at present altitudes of nearly 5300 feet at the east base of the Rocky Mountains in Alberta includes elements derived from the District of Keewatin at present altitudes of 1000 feet and less. The difference between these figures represents the minimum vertical lift. Actually the lift was even greater, because differential warping of the crust has since raised the Keewatin region relative to the Alberta region by an unknown amount.

Early in the study of the American drifts it was held by T. C. Chamberlin that the vertical component in the transport of drift results from localized oblique upward shear. This is undoubtedly true at the margin of an ice sheet where, as indicated in Chapter 2, obstruction to flow causes shearing of that kind. However, throughout the great mass of an ice sheet, movement consists of flow. The streamlines of flow at the base of the ice sheet curve upward toward the position of the most rapidly flowing ice (Fig. 2), and where flow is obstructed toward the terminus all streamlines are inclined upward. The movement is analogous to the slow forward and upward movement of water at the bottom of a basin or pocket in a river channel, or of the ice in a deep bedrock basin beneath a valley glacier. In the example from the Presidential Range cited above, the glacial boulders at the higher altitudes are confined to the stoss (north and west) slopes of the mountains. Apparently the load was dragged up the slopes by the basal ice rather than dropped into place from positions higher up in the ice.[16] Similar confinement of the drift to stoss slopes has been reported from several mountain districts and is probably general.

QUANTITATIVE ASPECTS OF GLACIAL EROSION

Abrasion and Quarrying Compared

Abrasion, the scour of rock by rock, is evident not only in the striated, polished, and fluted surfaces common to most glaciated regions but also in the vast quantities of fine sand and silt, "rock flour," visible in the streams that drain living glaciers and in the deposits of former glacial

[16] R. P. Goldthwait 1940.

streams. It is obvious that the grinding up of bedrock into fine particles is an important element in glacial erosion. On the other hand quarrying is evident in the steplike profiles of the floors of some glaciated valleys and of the lee slopes of glaciated hills and bosses, as well as in the stones and boulders that make up a small but conspicuous element in the deposits of glaciers. Therefore quarrying, like abrasion, is a large element in glacial erosion. Which has contributed the larger share to the sum total of rock eroded?

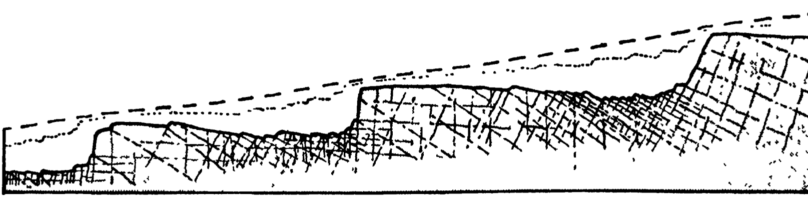


FIG. 17. Long profile and section of part of a glaciated valley. (After Matthes.) Surfaces smoothed by abrasion alternate with slopes steepened by quarrying, imparting to the valley a steplike profile. In this valley the steps are shown to be controlled by the unequal distribution of joints and other planes of weakness throughout the bedrock. The profile prior to glaciation is suggested by the dashed line. The dotted line suggests a profile made early in the process of glacial erosion.

To this question the answer must be that, by and large, and with local exceptions, quarrying has removed far more rock than abrasion. This is shown by the long profiles of steep glaciated valleys, which demonstrate greatest excavation where joints and other surfaces of weakness are most closely spaced (Fig. 17). It is shown also by the greater amount of erosion evident on lee (quarried) slopes than on stoss (abraded) slopes of hills in areas eroded by ice sheets. A quantitative study of the granite hills of eastern Massachusetts established the relation between glacial erosion and sheet structure.[17] This structure consists of concentric layering roughly parallel with the surface of the ground. The sheets become thicker and more nearly horizontal with depth, and they extend hundreds of feet below the surface. Now, although on the stoss sides of hills this structure (which is preglacial) is essentially parallel with the present surface, on their lee sides the sheeting is inclined more gently than the present surface (Fig. 18). Clearly the stoss slopes were mainly abraded while the lee slopes were mainly quarried. An empirical curve[18] showing the thickness of sheets relative to their depth below the surface

[17] Jahns 1943.
[18] Jahns 1943, fig. 16.

makes it possible to determine the approximate amount of erosion that has occurred at each place. Through the use of this curve it appears that abrasion took more off the sides of hills than off their stoss slopes. More important, it appears also that quarrying took several times more rock from the lee slopes than abrasion took from the stoss slopes. Erosion was greatest where joints cutting the sheets were most closely spaced.

This group of determinations is more exact than is generally possible, but the results are similar to those obtained elsewhere by less refined means. They show that in general quarrying is quantitatively more effective than abrasion.

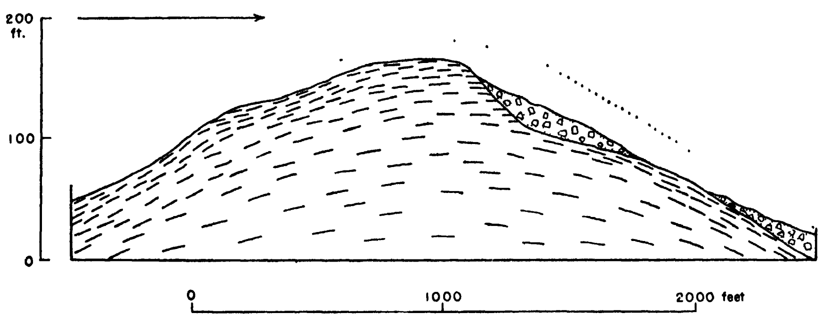


FIG. 18. Idealized section through glaciated hill of granite showing relation of present surface to sheet structure, and inferred preglacial surface (dotted line). The former ice sheet moved from left to right, accomplishing more quarrying of the lee slope than abrasion of the stoss slope, and depositing drift in places. (Slightly modified after Jahns 1943.)

This conclusion is in accord with theory. It takes much less work per unit volume to quarry out most kinds of rock than to wear rock down by abrasion. This is particularly true of hard, well-jointed rocks and is less conspicuously true of weak, poorly consolidated rocks.[19] In general we can say that glacial erosion will be expressed by quarrying as long as joints and similar surfaces are closely spaced and somewhat open and cut a surface that is subject to tensional stress as the ice moves across it.

However, it does not follow that, where ice-abraded surfaces are widespread, there abrasion exceeded quarrying. Such surfaces prove only that abrasion was the latest significant process to affect them; actually this process may have merely polished off an area that previously had been reduced mainly by quarrying. Over large areas this, in fact, was probably what happened, as is indicated hereafter.

[19] Clay-rich till is well consolidated in relation to glacial erosion and is virtually free of joints. Because of these facts it resists quarrying almost perfectly and, where glacially eroded, shows the effects of abrasion only.

Relation of Rate to Depth of Erosion

It was stated in Chapter 3 that in an ice sheet the most rapid flow occurs in the marginal part, and that in the central area movement is very slow. It was stated also that within the marginal region the rate of flow increases radially outward from the locus of maximum accumulation, a statement that we must modify to the extent that in the terminal zone both thinning and the accumulation of drift may constitute obstructions that reduce the rate of flow. Probably (although the matter has not been fully investigated) the rates of both abrasion and quarrying are roughly proportional to the rate of flow of the ice, assuming that rock debris is present in the ice to serve as tools. From this probability we can deduce[20] that beneath a large ice sheet erosion is considerable only near its margin; elsewhere the rate of erosion is slight. The application of this deduction to erosion by the Laurentide Ice Sheet in northern North America is discussed in the following section.

Even where the rate of flow is adequate, the depth to which erosion can occur is limited by the structure and physical condition of the rock material at and near the surface. Once a glacier has removed the weathered mantle and has worked down through the open-jointed bedrock beneath, the rate of quarrying becomes sharply reduced and continues to diminish as joints become tighter with depth. Quarrying is cut down to a minimum, the load of rock fragments in the base of the ice diminishes correspondingly, and even the relatively slow process of abrasion is thereby reduced. At this stage the bedrock surface has been worn down to a series of irregularly undulating bosses having striated and polished surfaces, especially on their stoss sides.[21] The Canadian Shield, and less conspicuously northwestern Finland — regions of hard rocks overrun by powerful ice sheets — exhibit thousands of square miles of this kind of surface.

The actual depth at which joints become tight enough to inhibit glacial erosion must vary with varying lithology, topography, and preglacial climate. The available direct evidence, some of it detailed hereafter, suggests that in a region of low relief this depth is no more than a few tens of feet at most. In mountain valleys glacially eroded rock basins with closures of many hundreds of feet prove that under favorable circumstances depth of erosion can be very great. In some mountain

[20] Coleman 1926, p. 17; Demorest 1943, p. 377.

[21] These bosses were recognized in the Alps as early as 1787 by Saussure (1786–1796), who described the rippled glistening effect produced by a whole series of them as "roches moutonnées," in fancied resemblance to contemporary wigs slicked down with mutton tallow. As Longwell (1933) remarked, this term has been widely misapplied and mistranslated. It is of doubtful value at best.

valleys it has certainly exceeded 2000 feet. Some of the factors that make this possible are (1) steep preglacial gradients, assuring high velocity to the flowing ice and making quarrying easier by minimizing the support that joint-controlled blocks of rock in place receive on their downstream sides; (2) concentration within narrow valleys of ice discharge from wide sectors of the snowshed, thus favoring high velocity of flow; (3) frost wedging of rocks above the snowfields along range crests, and valley widening by glacial erosion, providing increments of tools for abrasion even after quarrying beneath the glacier has been reduced to a small figure. Daly[22] noted that abrasion was predominant over quarrying in the deep glacial erosion of the low-gradient major valleys of the Rocky Mountains near the 49th parallel. If we grant that the abrasion evidenced was anything more than a final polish on an excavation made chiefly by quarrying, we may suppose that, after all the quarrying that was possible had been accomplished, the ice completed the excavation by the slow process of grinding down the surface.

Depth of Erosion: Regional and Local Examples

The glacial deposits on the plains of Germany, Poland, and Russia contain "immense quantities" of rock of Scandinavian origin. It has been calculated[23] that the volume of this drift is so great that were it removed it could be used to fill up the basin of the Baltic Sea and the basins of all the lakes in Scandinavia, and that enough would be left over to add a layer 80 feet thick to the surface of the entire Scandinavian peninsula. This suggests that glacial erosion in the Scandinavian Mountains was very great, especially since a large part of the rock glacially eroded from these mountains was transported not southeast to Germany and Russia, but westward into the Atlantic Ocean, where it is concealed from view. Kerr believed that glacial scour had deepened the large valleys of northern British Columbia and southern Alaska by "at least 2000 feet."[24] Reid concluded that the average load of glacial "rock flour" carried by the streams draining Muir Glacier in southern coastal Alaska corresponds to an annual loss of about 0.75 inch of rock from the entire area beneath the glacier.[25] At this rate of erosion 2000 feet of rock could be removed in about 30,000 years.

The figures cited are outstandingly large. But both the Scandinavian Mountains and the mountains of coastal Alaska offer optimum condi-

[22] Daly 1912, p. 581.
[23] A. M. Hansen 1894, p. 123.
[24] Kerr 1936, p. 681.
[25] H. F. Reid 1892, p. 51.

tions for glacial erosion. Both are high and steep, and both have maritime climates which during a glacial age would provide very heavy snowfall. Together with the similar mountains of Chile and the South Island of New Zealand these regions should be expected to exhibit profound glacial erosion.

The Canadian Shield is an example of the other extreme. Local evidence of slight depth of glacial erosion has been reported from many different districts. Unequivocal evidence[26] is furnished by the Flin Flon district, 400 miles southwest of Hudson Bay. Here marine sedimentary rocks of Ordovician age overlie pre-Cambrian rocks with an irregular unconformity having a relief of about 50 feet. Before the Pleistocene, the Ordovician cover had been stripped back, exposing an extensive area of the irregular sub-Ordovician surface. The whole region was then glaciated. But the stripped and glaciated surface has almost the same appearance and relief as the surface still unconformably covered by Ordovician strata. Evidently glaciation failed to modify more than the details of the relief. An intricate pre-Ordovician drainage pattern closely adjusted to weak belts in the pre-Cambrian rocks remains unaltered by the glaciation save for the excavation of shallow rock basins in some of the larger valleys.

Indeed the detailed adjustment of drainage to lithology, long antedating the glaciation and yet undestroyed by that event, is a feature that characterizes wide areas of the Canadian Shield. Widespread also in the Canadian Shield and in the resistant rocks of the New England region is the glacial removal of the weathered mantle. Out of thousands of exposures of bedrock examined in New Hampshire, only 46 included any chemically decomposed rock.[27] Before glaciation the weathered zone in both New England and the Canadian Shield was probably thin owing to slight permeability of the dominant rock types and to cold climate. Hence the removal of this mantle does not in itself argue deep glacial erosion. It is doubtful whether the average thickness of rock removed from the Canadian Shield by the recurring Laurentide Ice Sheet amounts to more than a very few tens of feet at most. It has been estimated that the somewhat similar area of Finland has lost 30 to 60 feet of rock through glacial erosion.[28]

Probably very little of this erosion took place during the times when Finland and the Canadian Shield lay beneath the central areas of the ice sheets. It occurred chiefly while the submarginal parts of the expanding glaciers flowed across those regions before the maximum of each glacial

[26] J. W. Ambrose, *unpublished.*
[27] J. W. Goldthwait and Kruger 1938.
[28] The late Erkki Mikkola, *unpublished.*

age. No doubt, also, glacial erosion was promoted by the cumulative results of weathering during the long interglacial ages.

By reference to the thickness of sheets in granite it was determined that 10 to 15 feet of rock had been glacially removed from the stoss slopes of hills in eastern Massachusetts and that variable amounts up to 100 feet had been quarried from lee slopes.[29]

The foregoing discussion leads to the generalization that in mountains with abundant snowfall glacial erosion has been deep, whereas, in regions of slight relief overspread by ice sheets, glacial erosion has been comparatively slight. Locally, however, topography plays an important part in diminishing or increasing erosion. Beneath an ice sheet a rock floor with a relief of many hundreds or thousands of feet and steep slopes as well interferes with and reduces the radial flow of the ice mass. This was true apparently of the part of the Cordilleran glacier complex in southern British Columbia and Washington, where despite the presence of ice some thousands of feet in thickness very little erosion occurred.[30] Yet in this same region an independent valley glacier, fed by abundant snowfall in the crest of the Cascade Mountains, excavated a rock basin (Lake Chelan) with a closure of 1040 feet.

Even in the lee of individual hills the local presence of preglacially weathered mantle, in many places overlain by undecomposed till, while the tops of the hills themselves are scraped clean, shows the protective effect of topography on a small scale.[31]

The most conspicuous example of slight glacial erosion throughout a wide area is the region extending from the crest of the Shickshock Mountains, the backbone of the Gaspe Peninsula, southward across Chaleur Bay through western New Brunswick. In this region of about 15,000 square miles glacial erosion is so slight that the terrain has been considered by more than one geologist to have escaped glaciation. Yet glaciated it was — probably repeatedly.[32] The slight erosion may have resulted from the fact that this region lies in the lee of a formidable barrier, the Shickshock Mountains, in places reaching altitudes of more than 4000 feet, lying at right angles to the direction of flow of the Laurentide Ice Sheet. As this glacier flowed southeastward from the highlands of Labrador and southeastern Quebec into the capacious trough of the lower St. Lawrence, the basal ice was diverted northeastward down the trough, scouring it vigorously, while the upper ice was barely able to clear the barrier. In consequence the rate of flow downstream from the

[29] Jahns 1943.
[30] Daly 1912, p. 581; Flint 1937, p. 228.
[31] Chalmers 1898.
[32] Flint, Demorest, and Washburn 1942.

barrier was effectively diminished and erosion was correspondingly inhibited.

At the opposite extreme is the valley trending parallel with the local radius of the ice sheet that enters it. The fact that the ice is thicker over the valley than over the adjacent uplands makes for increased erosion of the valley floor, and the lack of topographic obstruction enhances still further the capacity of the ice to excavate the valley. The result is valley deepening, often coupled with the excavation of rock basins in favorable segments of the valley floor. The floors of tributaries, excavated little or not at all, are likely to be left hanging above the deepened floor of the main valley.

Classic examples are the troughs of the Finger Lakes in central New York. These valleys were first cut by preglacial streams flowing north. They were widened, greatly deepened, and straightened by subglacial erosion when the Laurentide Ice Sheet repeatedly filled and overran them from the north. Dams of drift at their northern ends created the present lakes. The valleys of both Seneca and Cayuga lakes have tributaries that hang above the main valley floors by as much as 400 feet. The steep-sided trough of Seneca Lake is more than 1000 feet deep *below the lake level;* its bedrock floor has not been reached by deep borings. In this region conditions for ready glacial erosion were nearly optimum: the valley floors are excavated in shales that dip gently southward, so that they continually presented a hackly surface to the sole of the flowing ice.

Despite this deep erosion of the valleys, the tops of the adjacent uplands in some places still retain preglacially weathered mantle, showing that the ice at this altitude (1500 to 2000 feet above the excavated valley floors) was either too thin or too poor in basal load to accomplish much erosion.

Another example of exceptionally great glacial erosion controlled by topography is the Highlands segment of the Hudson River valley between Newburgh and Peekskill, New York.[33] Here the river follows a preglacial gorge through a massive highland 1400 to 1600 feet high. This highland formed a barrier transverse to the direction of flow of the Laurentide Ice Sheet. Through the gorge the bedrock floor of the Hudson valley is a rock basin 765 to 950 feet below sealevel, and tributary valleys hang above it. In contrast, through the lowlands both up- and down-stream from the gorge the rock floor of the valley is generally not lower than 300 feet below sealevel. The basin is attributed to increased glacial erosion caused by the more rapid flow required for the ice upstream to pass through the gorge.

The valley of the North Branch of the Susquehanna River between

[33] Cf. Berkey 1911, p. 81.

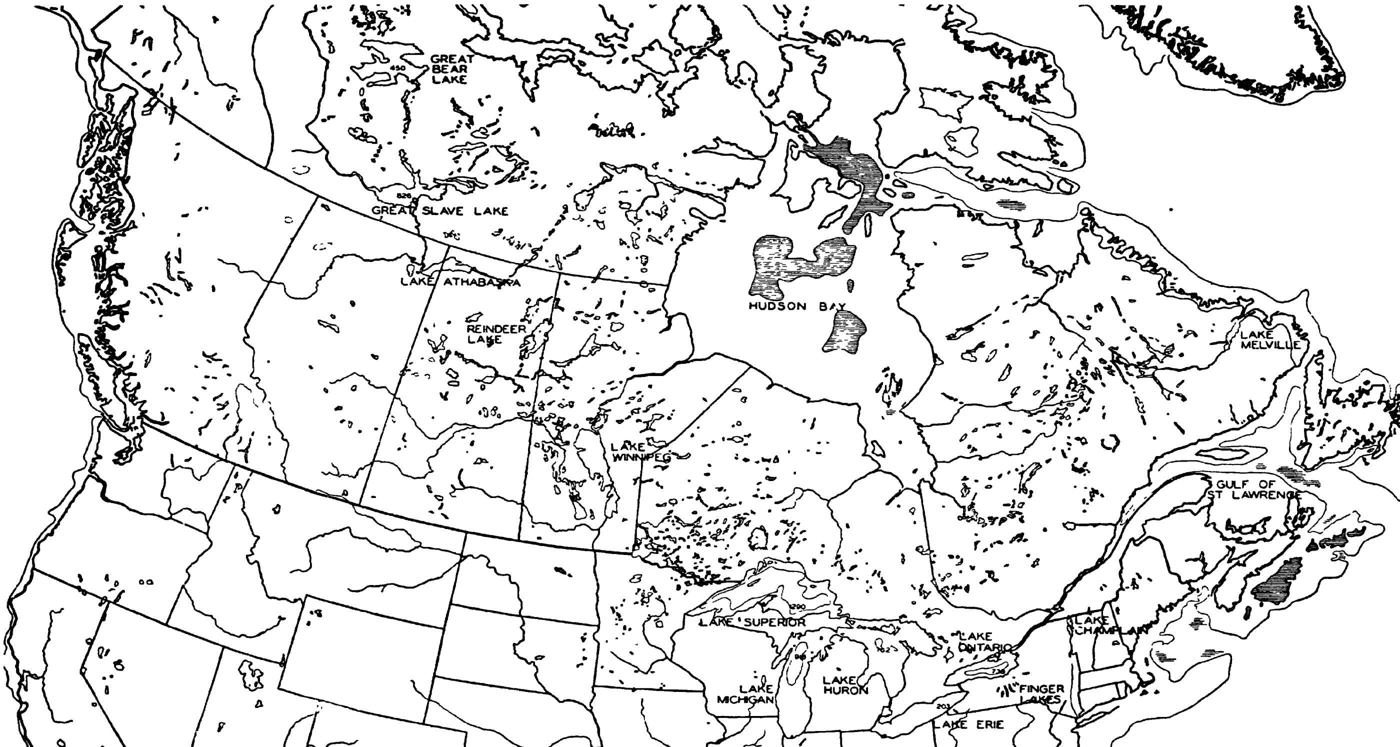


Fig. 19. Chief basins in North America, most of them of glacial origin. (They include not only rock basins but also basins made by dams of glacial drift and by crustal movements resulting directly from glaciation.) Basins in the continental shelf are shaded: form line on the sea floor is at 500 feet below sealevel. (Shepard 1937.)

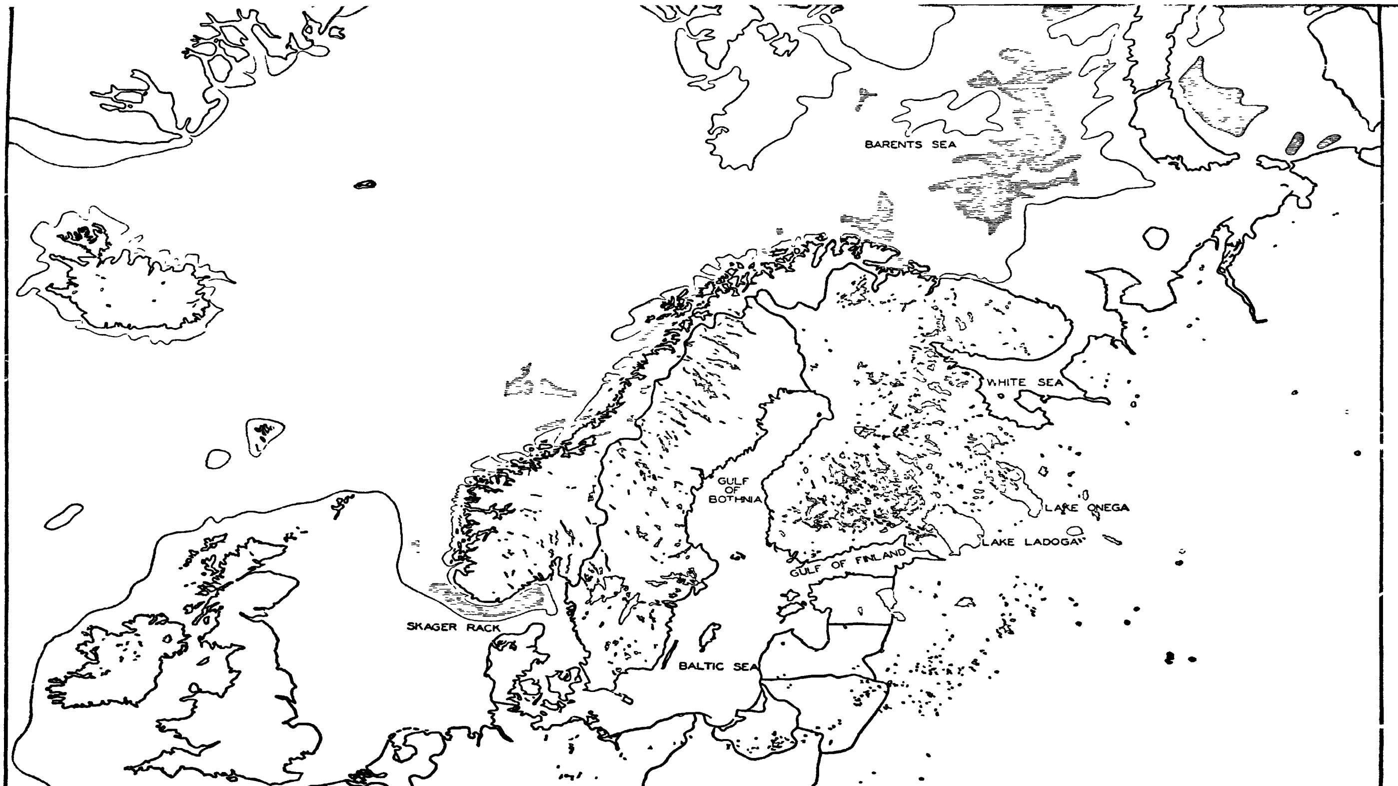


FIG. 20. Chief basins in northern Europe. Conventions are as in Fig. 19. (Shepard 1937.)

Pittston and Wilkes-Barre, Pennsylvania, consists of a series of rock basins with closures up to 225 feet. The valley of the Connecticut River in northern Connecticut and southern Massachusetts[34] has a long rock-floor basin with a closure of unknown amount. In both valleys the basins are believed to be a result of deep glacial excavation.

Many of the lakes in northern North America and Europe occupy rock basins excavated by ice sheets or valley glaciers in places where lithology or topography made erosion especially easy. It is probable that the Great Lakes of North America owe their basins chiefly to local glacial excavation of the floors of broad preglacial valleys. Figures 19 and 20 show the distribution of basins in North America and Europe believed to be of glacial origin.

It was stated in the preceding section that beneath a large ice sheet erosion should be considerable near (though not at) its margin and very slight beneath its great central area. This principle has been applied to the region covered by the Laurentide Ice Sheet in North America through the assertion that in that region three concentric zones exist. They are said to consist of an outer zone of predominant deposition (mainly the Great Lakes and Plains regions), an intermediate zone of predominant erosion (mainly the greater part of the Canadian Shield), and a central zone of little erosion (mainly the region east of Hudson Bay). Although the principle is probably valid, the application is only partly so. Unquestionably a broad marginal belt extending westward and northward from the eastern Great Lakes does have much more drift than the region north and east of it. But this belt coincides fairly closely with outcrops of comparatively weak rocks, soft Paleozoic and Mesozoic strata. Eastward from the Great Lakes, though still in the marginal belt, the drift becomes much thinner and less continuous, and this change coincides with a change in the outcropping rocks to much more resistant types. The intermediate zone of erosion coincides fairly closely with the zone of very strong rocks of the Canadian Shield. The only area within this zone where very thick drift has been reported[35] is the area immediately southwest of Hudson and James bays, where the outcropping rocks are soft Paleozoic strata. These facts suggest that the determining factors are lithologic rather than related to position with respect to the ice sheet. Finally the observations on which the supposed central zone of little or no erosion is inferred — surface rocks broken up by frost wedging rather than conspicuously eroded by glacier ice[36] — may indicate postglacial destruction of glaciated surfaces by frost wedging rather than lack of glacial erosion.

[34] Flint 1933.
[35] Dyer 1928.
[36] Cf. Low 1906, pp. 223–224.

Indeed the very complex history of the Laurentide Ice Sheet is opposed to any such simple view of the distribution of erosion. At one time or another the marginal part of the ice sheet swept over nearly the whole region covered by the ice at its maximum. Furthermore there were several different centers of outflow, which shifted their positions from time to time, and which should have given rise to very complex patterns of erosion and deposition. However, as lithology appears to be the dominant factor in controlling erosion and deposition, such patterns would be indistinct if recognizable at all.

The zonal concept of the distribution of glacial erosion seems to fit the Cordilleran region somewhat better than the Laurentide region. In British Columbia both the Coast Ranges and the Rocky Mountains bear the marks of much deeper and more powerful glacial erosion than do the lower mountains of the "Interior Plateau" region that separates them. This results partly from the fact that the preglacial topography with its steep narrow valleys favored deep erosion in the high mountains, whereas the valleys of the Plateau are more open. More important, it results from the much longer occupation of the mountains than of the Plateau by glacier ice. The glaciers formed in the mountains and widely invaded the intervening country only near the glacial maxima. The invading glaciers consisted of thick piedmont ice masses which at times may have coalesced into an ice sheet. During the occupation of the Plateau by widespread ice, movement was slow and over large areas erosion was slight. Indeed the fact of glaciation in the centrally placed Cariboo and Cassiar districts was questioned for a time[37] although detailed study[38] later established it beyond doubt. Other areas within this inter-range Plateau have been described as recording very slight glaciation although formerly overwhelmed by ice.[39]

Relative Rates of Glacial and Subaerial Erosion

Both glacial erosion and subaerial erosion are controlled as to rate by so many variable factors that any quantitative comparison of them can have only a very generalized significance. Both before and for some years after the turn of the nineteenth century a lively controversy among geologists centered on the question whether glacial erosion is more or less effective within a given time than subaerial erosion.[40] As has occurred

[37] J. B. Tyrrell 1919.

[38] Johnston 1926.

[39] N. F. G. Davis and Mathews 1944.

[40] Good accounts of the controversy, with summaries of some of the scores of papers written on this subject, are Bonney 1910, Carney 1909*b*, and Garwood 1932.

time and again in similar controversies the matter was complicated unnecessarily, and general agreement was delayed, by the fact that not all those concerned were considering the same aspect of the problem. Some were drawing inferences from localized zones of specially effective erosion while others based their arguments on the facts in areas of dominant deposition.

The amount of energy available for erosion through the descent of a given amount of precipitation from top to base of a given slope is the same whether the precipitation is in the form of water or in the form of ice. But, because of its slow rate of flow, ice fills a valley more fully than a stream of water of comparable discharge. In consequence it can erode a much larger area in the same time. Further, because of its great viscosity, its competence to carry large rock fragments is much greater than the competence of streams of water, which must slowly wear down large boulders to a size that can be pushed. In a single mountainous region the great turbidity of streams fed by glaciers compared with that of normal streams furnishes direct evidence that on the same slopes and on the same kinds of rocks valley glaciers are the more efficient agents of erosion.[41]

Although this generalization is true of valley glaciers flowing swiftly on steep gradients, it is gradually being recognized that ice sheets, even in their submarginal areas, may be only temporarily effective erosive agents. Owing to the gradual transition from rapid erosion of the superficial zone to very slow erosion of the deeper zone of fresh rock with tight joints, it is probable that the diminishing rate of glacial erosion in a region of low relief first exceeds but later fails to equal the rate at which subaerial agencies might be expected to reduce the same region. The ice sheet prevents weathering of the rocks beneath it. In its later phases it is, relatively speaking, a protective rather than an erosive agent. This condition has been inferred[42] both for the Greenland Ice Sheet, which today is delivering very little drift to its western margin, and for the Antarctic Ice Sheet,[43] which is moving very slowly and is carrying an apparently negligible load of rock waste.

The landscape features sculptured by various types of glaciers are discussed in the following chapter.

[41] T. C. Chamberlin and R. T. Chamberlin (1911, p. 207) and von Engeln (1911, p. 142) present good detailed comparisons. However, as most glaciers today are shrinking, the turbidity of their meltwaters is greater than it would be, were the glaciers expanding or in equilibrium.

[42] Demorest 1937, p. 46.

[43] Gould 1940, p. 866.

Chapter 6

GLACIALLY SCULPTURED LANDSCAPES

THE GLACIATED VALLEY

Long Profile

Virtually no valley is excavated in the first instance by glacier ice. Nearly every glaciated valley is a pre-existing valley, usually a stream valley, that has been occupied and more or less remodeled by glaciers. Hence the best way to study the erosive effects of glaciers on valleys is to compare glaciated valleys with stream valleys, noting the significant differences. In general the glaciated valley has a long profile that is steeper in the headward part than that of the nonglaciated valley. The long profile is likely to have, moreover, a conspicuous steplike form in its headward part. Further, both between the steps, and farther down the valley where steps are rare or absent, there are likely to be rock basins.

For many years the development and maintenance of the steep risers of the valley steps were believed to be the result of the process described by Willard Johnson in 1904.[1] This consisted of the melting of ice in great transverse crevasses extending down to the bedrock floor, percolation of the meltwater into joints and other openings in the rock, and springing-out of blocks of rock by the force of expansion of the ice formed by refreezing of the water. A transverse bedrock wall would thus take form directly along the line of each crevasse. Frost wedging, as this springing-out and dislodgment process is sometimes called, is common at high altitudes wherever suitably jointed rocks and water are exposed to a great daily range of temperatures embracing the freezing point. But it can not be the main factor in causing valley steps beneath glaciers for the simple reason that few crevasses reach down to the rock floor beneath the ice. Shortly before his death Johnson himself abandoned his hypothesis.[2] A more satisfactory explanation[3] is based on the fact that steps have been observed to coincide with very poorly jointed parts of the bedrock. It attributes the steps to the resistance of massive, poorly jointed rock to erosion compared with rapid quarrying of well-jointed rock immediately

[1] W. D. Johnson 1904.
[2] Matthes 1930, p. 94.
[3] Matthes 1930, pp. 89–97.

down the valley (Fig. 17). This appears to be the origin of most valley steps, although it is likely that some steps are formed in other ways.[4] Notable among these is the steepening of slope caused by increased erosion where large tributaries join the main glacier.

Some rock basins, too, have this origin, for they are developed in transverse belts of rock exceptionally well provided with planes of weakness. Other basins record the excavation accompanying the narrowing of the cross-sectional area of a valley where the glacier was forced to flow through specially resistant rock. Still other basins may have developed immediately below the points of entrance of tributary glaciers, where the ice discharge was noticeably increased.[5]

Many glaciated valleys, among them a large number of Norwegian fiords, have in their outer or downstream parts long gently sloping rock basins. The closures of some of these basins exceed 3000 feet. The basins occur at so nearly the same segment in each valley that a common explanation of their origin seems probable. Brögger[6] attributed them to this factor: the rate of flow of a glacier increases downstream as far as the névé line. From the névé line downstream the flow decreases because excess of ablation over supply thins the ice more and more. In consequence the bedrock floor should be eroded to form a basin at the névé line and through some distance upstream from it. Many basins in the floors of valleys at or near their debouchures are more simply explained as the result of decrease of erosive capacity with decrease of velocity at the point where a narrow valley glacier merges into a broad, thick piedmont glacier. Even more simply, some may mark the places where flotation in the sea ended the erosive capacity of valley glaciers.

The gentle rise that separates a basin from the gently sloping valley farther downstream has been appropriately referred to as a *threshold*. However, some thresholds, especially those known only as a result of submarine soundings in fiords, may consist wholly or partly of moraines and other glacial deposits rather than of bedrock.

It is worth noting that both steps and basins are known to be present in the floors of stream valleys that have never been glaciated, although they are far less common in such valleys than in glaciated troughs. At any rate steps and basins are not in themselves criteria of glacial action.

Cross Profile

Glacial alteration of a stream valley includes both deepening and widening. In some valleys the volume of rock excavated by deepening

[4] Cotton (1941*a*) gives a good summary of ideas concerning these steps.
[5] Penck 1905.
[6] A. M. Hansen 1894, p. 124.

exceeds that excavated by widening, but in many valleys widening has been the more important process.[7] These changes result, in most valleys, in a pronounced U-shaped cross profile. To be sure, some stream valleys have a U shape, but it develops very slowly and the sideslopes are graded, whereas the glaciated U valley develops rapidly and the sideslopes are not graded. Most glaciated U valleys are made by alteration of youthful V valleys cut by streams and mass-wasting. Before glaciation, active downcutting was taking place only in the narrow valley floor where the stream was flowing; the sideslopes were being eroded chiefly by mass-wasting. If the valley was then filled with flowing ice, erosion became active on the sides as well as on the bottom, in fact around the entire ice-covered perimeter. Glacial erosion tends to remodel a valley toward a semicircular cross section because that shape has the least area in proportion to the volume of ice flowing through the valley and therefore offers the least frictional resistance to the flowing glacier.

Some glaciated valleys, particularly some of those in the Alps, have composite cross profiles, that is, profiles consisting of a narrow U set within a higher, broader U.[8] Several different explanations for this feature have been offered. Of them, two appear probable. Penck and Brückner[9] held these valleys to be preglacial stream-cut two-story valleys later modified by glaciation without obliteration of the preglacial two-story character. On the other hand Visser[10] explained them as preglacial stream valleys later converted by glaciation into U valleys, still later subjected to interglacial rejuvenation by stream erosion, and finally slightly modified by renewed glaciation. Very likely both explanations are applicable in different mountain regions.

Hanging Tributary Valleys

The deepening or widening of a main valley by glacial erosion, at a rate more rapid than the rate at which the tributary valleys are cut (regardless of whether or not the tributaries are filled with glacier ice) leaves the tributary, at its junction with the main, higher than the main, that is, "hanging" above it. Upon deglaciation, the stream in the hanging tributary falls or cascades down to the main valley floor. The vertical drop may amount to several hundred feet. Even where all tributaries carry glaciers, the normal relationship of their floors to the floor of the main valley should be "hanging," even though the upper surfaces of

[7] Cf. Matthes 1930; W. O. Crosby 1928.
[8] Cotton 1941*b*.
[9] Penck and Brückner 1909, pp. 288, 305, 376, 617, 837.
[10] Visser 1938, p. 140.

tributary and main are accordant at the point of junction, because the rate of glacial deepening of the valley floor depends very largely on the thickness of the ice.[11]

That deepening and not merely widening of the main valley is responsible for some hanging tributaries is said to be shown by the fact that long profiles of the tributaries, projected into the main, fall well above the floor of the main at its center line. If it can be shown that this lack of accordance is the result of glacial erosion rather than preglacial stream erosion, the case is proved. In view of the abundant evidence of great deepening of valleys by glaciers, however, the matter is academic.

The reservation mentioned carries with it the statement that hanging tributaries are not criteria of glaciation. They have been observed in nonglaciated highlands[12] where they can result from any process, such as crustal warping, that rejuvenates the stream in the main valley while affecting the tributaries less or not at all. It should be added that hanging tributaries even though of glacial origin do not constitute evidence of erosion by valley glaciers as compared with ice sheets. If in an extensive ice sheet the ice over a main valley is thicker and less impeded by topographic barriers than the ice over the tributaries, the tributaries can be left hanging. Examples are the hanging tributaries of the Finger Lakes valleys mentioned in a foregoing section.

Fiords[13]

Many of the strongly glaciated valleys of high-standing coasts underlain by resistant rocks in high latitudes[14] are partly submerged and constitute fiords. A fiord is merely a segment of a glaciated trough partly filled by an arm of the sea. It differs from other strongly glaciated valleys only in the fact of submergence.

In the past, extreme views on the origin of fiords have been held. These valleys have been regarded as tectonic, controlled by faults, fractures, and joints, later somewhat modified by streams and glaciers.[15] Fiords have been thought of also as almost wholly the product of glacial erosion, and by others as deep stream valleys only lightly glaciated. All these factors are present in fiords, but the relative importance of each varies from region to region. It is no more possible to generalize about the origin of fiords than about the origin of other types of valleys, beyond

[11] Penck 1905, p. 7.
[12] Matthes 1930.
[13] See Ahlmann 1919; Hubbard 1934; Peacock 1935.
[14] Norway, Scotland, Iceland, Greenland, Labrador, many Arctic islands, Alaska, British Columbia, Chile, Patagonia, and New Zealand.
[15] Cf. Gregory 1913.

the statement that all fiords were stream valleys prior to glaciation and all have been partly submerged. Some of these valleys were conspicuously controlled by rock structures; others are independent of them. Some were only moderately glaciated, whereas others were profoundly widened, deepened, and basined by glacial erosion.

The glaciation of some valleys took place while their floors stood above sealevel, submergence taking place later. Other valleys received their glacial deepening largely below sealevel. A glacier flowing down a valley cut in a mountainous coast partly submerged, and advancing into tidewater, excludes the sea and exerts undiminished pressure on its floor until flotation occurs. As the density of glacier ice is roughly 0.9, a glacier 3000 feet thick would continue to erode its floor even when submerged to a depth of nearly 2700 feet. Basins with closures of more than 3000 feet have been revealed by soundings. Probably the majority of glacial rock basins in fiords were excavated well below sealevel, as it is unlikely that postglacial submergence of most fiord regions has amounted to more than a very few hundred feet. In contrast note these great depths:[16]

GREATEST KNOWN FIORD DEPTHS[17]

In British Columbia	2574 ft. (in Finlayson Channel)
In Alaska	2898 ft. (in outer part of Chatham Strait)
In Europe	4000 ft. (in Sogne Fjord, Norway)
In World	4250 ft. (in Messier Channel, Patagonia)

THE CIRQUE

MECHANICS OF SCULPTURE[18]

The heads of many glaciated valleys other than valleys formerly occupied by outlet glaciers are shaped like a theater,[19] or half a bowl. Like bowls these valley heads vary in form: their floors are narrow to broad, their sideslopes moderately steep to nearly vertical. They are widely known as cirques,[20] and are believed to be sculptured by a combination of nivation and glacial scour. Nivation, the work done by the refreezing of snow meltwater, comes first and works like this.[21] A residual snowbank in a niche or slight depression undergoes daytime

[16] Peacock 1935, p. 669.

[17] Figures given are water depths; depths to bedrock may be greater.

[18] See Ahlmann 1919, pp. 175–178.

[19] Frequently miscalled amphitheater, which is like two halves of a bowl put back together, or a whole bowl.

[20] British *corrie* or *cwm*, German *Kar*; Norwegian *Botn*.

[21] Matthes (1900, p. 183) first recognized the process and coined the term. See also Bowman 1916, pp. 285–294.

melting in summer. The meltwater penetrates the underlying mantle and crevices in the bedrock beneath the whole area of the snowbank, freezes at night, expands, and wedges out small rock fragments. These are moved downslope by solifluction and by small rills. The effect is to deepen the depression, flatten its floor, steepen its sidewalls, and shape it so that its ground plan approaches the semicircular (Fig. 21). All this can and does take place beneath snowbanks too thin to form ice and flow.[22] The only movement in the snow is the movement of compaction.

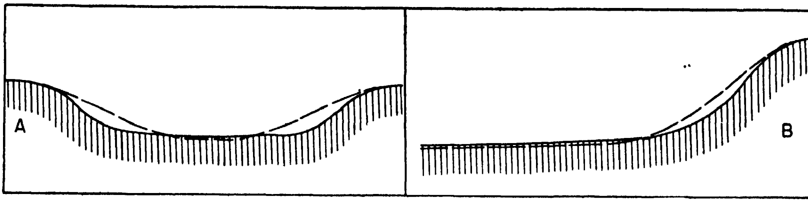


FIG. 21. Cross profile (*A*) and long profile (*B*) of head of shallow valley modified by nivation. Broken lines represent profiles before nivation began. Dotted lines represent upper surface of former bank of névé.

In areas where there are perennial snowbanks but no glaciers small cirques are fashioned solely by the process of nivation. Cirques of this kind[23] are evidently not in themselves an indication of glaciation.

But as soon as the snowbank becomes thick enough to flow, that is, becomes a very small glacier, solifluction and rills cease to be the only agencies in the removal of frost-wedged rock fragments. They are accompanied by the slow downslope transfer of rock debris by flowing ice, and therewith true glacial erosion begins. Evidence of this flow consists of glacial abrasion of the floors and lower sidewalls of cirques.[24] A cirque enlarged in this manner is a true glacial cirque. As some cirques reach depths of many hundreds of feet it is not easy to understand how frost wedging can occur beneath comparable thicknesses of ice and snow. At depths down to the lower limit of crevassing, perhaps 150 feet, frost wedging can occur as a result of fluctuations of temperature through the temperature-melting point. At greater depths frost wedging may result from melting and refreezing of ice as the pressure fluctuates

[22] See Hobbs 1910*b*, figs. 3, 4.

[23] Termed *nivation cirques* by R. J. Russell (1933, p. 936). Essentially similar forms are sculptured beneath present-day snowbanks in the loess hills of the Palouse region of southeastern Washington and southwestern Idaho. (See Rockie 1934; Kirkham and others 1931, p. 207). Pieces of rock, riven from the inclosing walls of nivation cirques cut in rock, slide and roll down the surface of the snowbank and accumulate at its toe in heaps resembling small end moraines (Daly 1912, p. 593).

[24] Bowman 1916

through the pressure-melting point with changes in accumulation and flow within the glacier.

Size, Form, and Orientation

A cirque may be of any size or proportion, depending on such factors as preglacial form of the valley head, regimen and duration of the glacier, and, above all, make-up of the rock from which it is cut. Other factors neglected, the volume (most obviously, the length) of a glacier is roughly proportional to the capacity of the cirque or cirques at its head. Single cirques form the heads of small glaciers only. Long valley glaciers head not in single cirques but in many small tributaries each with its own cirque, or in a single broad upland basin, formed by the coalescence of many adjacent cirques.

Many cirques are deeper near their headwalls than in their outer parts, with the result that their floors are rock basins, some of them containing lakes. Such cirques have, as Willard Johnson put it, a down-at-the-heel appearance. This form seems to be another indication that the ice even at the very head of the glacier does flow. Apparently the flowing ice quarries out blocks of bedrock and transports them upward as well as outward, creating a basin where rock structure is favorable.

In the northern hemisphere cirques are generally larger and more numerous on north and east slopes than on south and west slopes, just as small glaciers and snowbanks are today. This distribution is found even in the regions where the snow-bringing winds are westerly. Evidently fallen snow is swept from the summits into the more protected places on the lee slopes, where it builds perennial snowbanks and glaciers comparatively well protected from the solar heat that touches the south sides of the summits. Protection from exposure to the Sun, especially in summer, is more effective than the relative accumulation of snow in controlling the orientation of cirques.

Relation to Regional Snowline

The relation between the altitudes of cirque floors and the regional snowline is significant because it forms the basis of the principal means of estimating the altitudes at which the regional snowline stood during the glacial ages, as explained in Chapter 20. It is evident that within a given set of climatic conditions cirques can be excavated at or above the regional snowline but, save in exceptionally well-protected places, not far below it. Hence the altitudes of the cirque floors on a highland formerly glaciated are a rough indication of the altitude of the regional snowline at the time when the cirques were excavated. Very roughly,

within any restricted highland area the cirque floors lie within a fairly distinct range of altitudes. This fact strengthens the inference that their positions are strongly influenced by the regional snowline. However, there are many variations caused by details of topography, kind of rock, and exposure to the Sun. Furthermore, whereas in districts with light snowfall cirques appear to be formed fairly close to the regional snowline, the heavier the snowfall the farther above the snowline they tend to be. Various empirical formulas have been attempted, based on the lengths of the glaciers heading in the cirques.

All these calculations are more or less inaccurate because of local irregularities, but the fact is that if the irregularities are smoothed out in a calculation of regional scope the results are broadly consistent. Hence perhaps the chief significance of cirques is their use in approximating the regional snowlines of the glacial ages, from which the subsequent broad climatic changes can be inferred.

It should be emphasized that no former snowline can be measured directly. What can be measured is a feature that might be appropriately called the *level of cirque excavation*.[25]

In most mountain areas only a single set of cirques is found. That is to say, there is very little evidence of cirques formed at different altitudes in successive glacial ages. In view of the indubitable evidence of successive glaciations in plains regions, we can justly infer that, although most existing cirques were occupied during the latest great glaciation, cirques nevertheless existed during earlier glaciations. Probably some of the earlier-formed cirques were entirely destroyed by rapid frostwork before the advent of a subsequent glaciation. But it is probable that most cirques were reoccupied and refurbished by successive generations of glaciers. As they were deepened and steepened they became increasingly protective of the snow they contained and so became with increasing certainty the sites of later snow accumulation. According to this view the existing cirques are composite products of all the glaciations and are related to a composite Pleistocene regional snowline rather than to the snowline of any particular glacial age alone.

SIGNIFICANCE OF GLACIATED VALLEY FORM

None of the features of a glaciated valley is in itself diagnostic of glaciation. Each individual feature is found in regions never glaciated. The U-shaped cross profile is seen in many mature valleys sculptured by streams and mass-wasting. The floors of river valleys are commonly

[25] This level approximates what has been called the *orographic* (that is, *topographic*) snowline. See a discussion in Ray 1940, p. 1910.

marked with basins. Hanging tributaries are created through deepening of a main valley by streams alone. Valley heads with perfect cirque shapes are numerous in regions like the Grand Canyon of the Colorado River, far removed from glaciation. But the combination of these features has never been reported from a valley not glaciated, for since each one, if not glacial, is the product of a different cause, the occurrence of all of them in a single nonglaciated valley is highly improbable. It is the weight of evidence rather than the presence of a single diagnostic feature that justifies the inference of glaciation.

ALPINE SCULPTURE[26]

The most compact way in which to show the sculpture that takes place when valley glaciers form in a mountain region is by a series of sketches of an ideal landscape undergoing sculpture of this kind. Figure 22*A* represents an ideal mountain region before glaciation. As the climate becomes colder snow persists through the summer, forming banks of névé. Nivation begins to enlarge the heads of the pre-existing valleys. In time the masses of névé become thick enough to flow, and thus form glaciers. This condition is shown in Fig. 22*B*. The glaciers are most numerous on the right-hand sides of the mountains because these sides face away from the Sun. Cirques begin to develop at the glacier heads. The meltwater from the ice carries the rock waste down the steep tributary valleys and deposits it in the less steeply sloping main valley.

Expansion of the glaciers at length causes them to coalesce, forming a large trunk valley glacier in the principal valley (Fig. 22*C*). The glacier-occupied valleys are widened and deepened, the tributary valleys are left hanging above them, and the spurs between the tributaries are blunted and beveled as the larger glaciers grind past them. Enlargement of the valleys tends to straighten them.

While the valleys are being enlarged the crests of the mountains are sharpened by frost wedging along joints. The continued growth of cirques on opposite sides of a crest eventually reduces the crest to a knife-edged form (an *arête,* so named by climbers in the Alps) kept sharp by frost wedging. As a result the mountain range develops a sharp main ridge with sharp lateral spurs.

Where two cirques enlarging toward each other cut through the ridge that separates them, a sharp-edged gap (a *col* in Alpine climbers' terminology) with a smoothly curved vertical profile results. Many Alpine passes have this origin. Some of them have lost their sharp edges as a

[26] See W. M. Davis 1906; Hobbs 1910*b*, Ahlmann 1919.

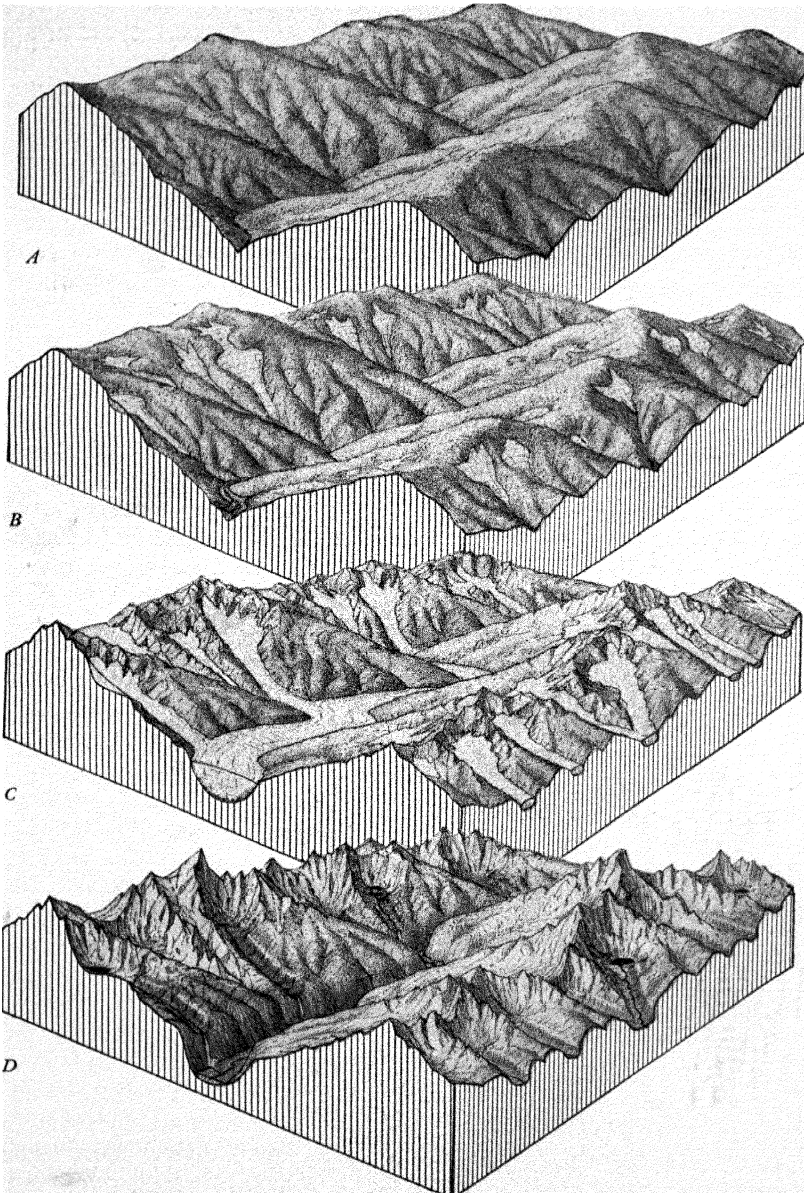


FIG. 22. Alpine sculpture (Longwell, Knopf, and Flint).

A. A mountainous area before glaciation.

B. Growth of small glaciers.

C. Development of a network of valley glaciers.

D. The same area after deglaciation, showing glaciated troughs, rock basins, faceted spurs, hanging tributary valleys, cirques, arêtes, cols, and horns.

result of glacial scour at times when glaciers expanded and flowed through them, converting them into smoothed U troughs.

Three or more cirques gnawing inward against a single high part of the mountain crest can sculpture the high part into a pyramid (called a ***horn*** by climbers) with several facets, each facet being the headwall of one of the cirques. These pyramidal peaks rising above jagged crests are the dominating feature of mountain ranges sculptured by Alpine glaciers.

All these features appear in Fig. *22D,* showing the appearance of the mountains after the climate has become warmer and the glaciers have disappeared.

Because the sequence of land forms just described is well developed in the Alps, where the study of glacial erosion began, these forms are often referred to as ***Alpine***. In Alpine sculpture the details of the valleys are the work of glacial erosion, but the equally important details of the cirques and higher spurs and ridges are largely the work of frost wedging and other processes of mass-wasting.

Alpine sculpture is characteristic of mountainous areas occupied now or formerly by a network of valley glaciers that never grew thick enough to bury or nearly to bury the ridges and higher peaks. It also characterizes mountains formerly buried beneath glacier ice but later occupied for so long by a network of valley glaciers that rapid frost wedging of the higher ridges sharpened them and recreated the Alpine forms. Alpine sculpture is therefore not in itself evidence that the mountains were never buried beneath glacier ice.

MOUNTAIN-ICE-SHEET SCULPTURE[27]

In regions of heavy snowfall and cool maritime summers the system of valley glaciers that develops during the earlier part of a glacial age is only a preliminary phase of glacial development. The individual glaciers enlarge, thicken, and coalesce over the spurs and ridges that first separated them. In time they form a continuous or nearly continuous cover beneath which, except perhaps for scattered nunataks marking the positions of the highest peaks, both valleys and ridges lie buried (Fig. *23A*). This kind of cover, which has been described as a ***mountain ice sheet,*** formerly existed over large regions in British Columbia and Alaska, over parts of the southern Andes, and probably over some of the mountainous parts of northeastern Siberia.

The mountain ice sheet, moving under great pressure, rapidly modifies the delicate detail of spur and ridge wrought by mass-wasting under

[27] Kerr 1934, 1936; N. F. G. Davis and Mathews 1944.

the earlier or Alpine conditions. Horns and ragged arêtes are ground down to make domelike peaks and rounded ridges, and sharp cols are smoothed, deepened, and broadened, in some places linking in a continuous trough two valleys whose heads were formerly separated from each other by a wall-like col. The intensity of sculpture of all these features varies from place to place beneath the mountain ice sheet accord-

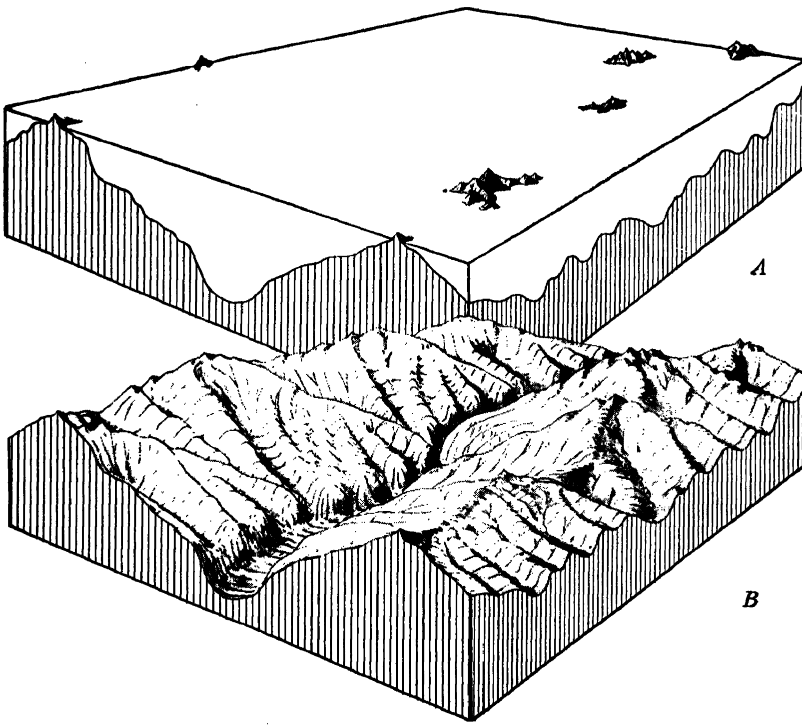


Fig. 23. Mountain-ice-sheet sculpture.

A. The area of Fig. 22, *D*, buried beneath a mountain ice sheet.

B. The same area after rapid deglaciation, showing smoothing and rounding of the formerly buried surface, contrasted with the reduced and frost-sharpened former nunataks.

ing to the distribution of snowfall on its upper surface and according to the form of the rock surface beneath it. If, as in the interior of British Columbia, the ice sheet fills a somewhat basinlike region having only a few narrow outlets, the intensity of erosion will be less than on a surface that does not confine the spread of the ice.

As the climate becomes warmer and the mountain ice sheet shrinks, thinning brings peaks, ridges, and spurs once more above the glacier surface (Fig. 23*B*). Small valley glaciers re-form in high protected places,

frost wedging sets in, and rounded surfaces are vigorously sharpened. So rapidly does this take place that perhaps, even before the last remnants of the ice sheet as such have vanished, Alpine sculpture has destroyed the smoothed and rounded summit surfaces by which the former presence of the ice sheet is recognized. It is probable that some Alpine regions were formerly covered by mountain ice sheets of which no distinct traces remain.

SCULPTURE BY PIEDMONT GLACIERS AND LOWLAND ICE SHEETS

There seems to be no essential difference between the sculpture to be expected from a large piedmont glacier and the sculpture wrought by an ice sheet upon a lowland. For the sake of completeness, however, these two types of glaciers must be grouped together because most if not all existing or former lowland ice sheets grew through the building up and expansion of piedmont glaciers, which in turn developed from the expansion and coalescence of many valley glaciers. The submarginal zone is the area of most active erosion, and outward beyond it, beneath the marginal zone, the bulk of the eroded rock waste is deposited. Where erosion is exceptionally strong the topography takes on a stoss-and-lee character. Rock basins are excavated at favorable places in the floors of valleys, and valleys are blocked by glacial deposits so as to form additional basins. Many of the sculptured details developed during the expansion of the glacier are buried beneath glacial deposits during the later shrinkage and so are concealed from view.

Chapter 7

GLACIAL DRIFT I: TILL; MORAINES

ORIGIN AND MEANING OF THE TERM DRIFT

Long before the glacial theory was announced, the earth and stones spread over large parts of the surface of Britain were recognized as deposits brought from elsewhere. They were thought of as having been "drifted" in by water or by floating ice, and so, quite naturally, they came to be called *drift.* Although rooted in a concept we know to be in error (except for the drift on the deep-sea floor which is a genuine iceberg deposit) this name was so well established by the time the real origin of the deposit was widely recognized that it became a part of the standard glacial vocabulary. As used today the term *glacial drift*[1] embraces all rock material in transport by glacier ice, all deposits made by glacier ice, and all deposits predominantly of glacial origin made in the sea or in bodies of glacial meltwater, whether rafted in icebergs or transported in the water itself. It includes till, stratified drift, and scattered rock fragments.

Thus interpreted, this material covers large parts (though by no means all) of the glaciated regions of the world and extends beyond them along stream valleys and on the floors of lakes and the sea, wherever water could carry drift away from the margins of the glaciers themselves.

THE DRIFT IN TERMS OF SEDIMENTS

THE SEQUENCE: TILL → STRATIFIED DRIFT

The drift can be classified in two different ways: in terms of sediments, specifically the arrangement of the pieces of rock of which it is composed; or in terms of its form or topographic expression. Unfortunately for simplicity of discussion both classifications have to be used despite the fact that they are only partly related to each other.

Geologists early subdivided the drift into two supposedly distinct kinds: till or nonstratified drift, and stratified drift. Thus they recognized a distinction that is fundamental though it is not universal. Till was soon

[1] The short form, *drift*, is used in this book even though Webster defines *drift* so as to include nonglacial deposits.

seen to be a direct glacial deposit, for its content of a wide range of grain sizes and the lack of obvious arrangement of its component particles show that the selective activity of water had played a minimum part in its deposition. In contrast stratified drift, as its name implies, is distinctly sorted according to size and weight of its component fragments, and thereby indicates that a fluid medium far less viscous than glacier ice — in other words, water or air — was responsible for its deposition.[2]

Till, like *drift,* is a term that long antedates the glacial theory. It is a Scottish word, used by generations of Scots countryfolk to describe "a kind of coarse, obdurate land," the soil developed on the stony clay that covers much of northern Britain. The earliest detailed areal glacial studies published in Britain were Scottish.[3] Hence the Scots term came into wide use rather than the English term *boulder clay.* This is fortunate because *boulder clay* is not a good designation for the whole range of deposits we know as till. It is not good because some till contains no boulders, some contains little or no clay, and some (though perhaps not very much) contains neither boulders nor clay, but only silt, sand, and small stones.

The more the drift is studied the clearer it becomes that there is no sharp dividing line between till and stratified drift, but that one grades into the other. And the more living glaciers are studied the more easily is it seen why the gradation exists. Meltwater is present far back in the terminal zone of the glacier, on, in, and under the ice, and this water can flush away some of the finer-grained rock material during or even before deposition by the ice. Sometimes the finer sedimentary fractions are entirely removed before deposition of the residuum takes place. In others pockets of stratified sediment accumulate in glacial pools and are incorporated in a mass of nonstratified sediment. Yet the whole is till.

Because of this we regard till as glacial drift dominantly nonsorted according to grain size. It is the non-size-sorted end member of a series whose opposite end member is well-sized-sorted stratified drift. Ideally till is formed without the cooperation of water, but actually size sorting is present to an indefinite degree in deposits to which the term is applied.

LITHOLOGY AND THICKNESS

The drift, whether stratified or not, consists predominantly of rock material that was fresh and undecomposed before it was deposited.

[2] Actually *washed drift* or the old-fashioned term *modified drift* would be better than *stratified drift* because it is more comprehensive. A great deal of drift has had the fine components washed out of it by water without having been actually stratified. But it is hardly likely that any attempt to supersede the well-established term would be successful.

[3] Cf Archibald Geikie 1863

Minerals like the hornblendes, micas, and plagioclase feldspars, notably susceptible to chemical decay, are conspicuous in drifts derived from bedrocks that contain these minerals. Most of the rock fragments in the drift are mechanically broken or abraded. All this means that for the most part the glaciers were eating into fresh rock rather than altered mantle. Chemical freshness characterizes the earlier glacial deposits as well as the later ones. The inference is plain that decomposed mantle did not contribute an important fraction of any of the drifts; the ice early got down below the weathered zone into the firmer rock beneath.

The thickness of the drift is so variable, and is dependent on so many factors, that figures have no great significance. It has been suggested that the average thickness of the drift in the Great Lakes region as a whole may be 40 feet.[4] More locally, the thickness in southeastern Wisconsin has been estimated at 45 feet,[5] in Illinois at 115 feet,[6] and in Iowa at 150 to 200 feet.[7] In central Ohio[8] more exact figures are at hand because 2800 well logs from an area of 4500 square miles have been compiled. The results are these:

	Feet
Average thickness of glacial deposits throughout entire area	95
Average thickness over buried uplands	51
Average thickness in buried valleys	205
Greatest thickness penetrated	763

This drift undoubtedly includes two or more glacial stages. In the New England States where the rocks are more resistant to erosion and where favorably oriented drainage exported a greater proportion of the drift from the region while the ice was present than was possible farther west, the average is no more than 15 to 20 feet. Where local traps for sediment were formed, stratified drift was built up to great thicknesses. Borings reveal thicknesses of 1100 to 1300 feet in the Spokane Valley in Idaho and Washington,[9] and of 1080 feet at Watkins, New York, in the deep valley of Seneca Lake.[10]

There is very little relation between the thickness of a pile of drift and the time it took to be deposited. The Sefström Glacier in Spitsbergen built a pile of till 100 feet thick in less than ten years.

[4] Quirke 1925, p. 394. See a drift-thickness map of Indiana in Leverett and Taylor 1915, pl. 4.
[5] Alden 1918, p. 151.
[6] Leverett 1899, pp. 542–549.
[7] Kay and Apfel 1929, pp. 181, 256.
[8] Ver Steeg 1933.
[9] Flint 1936, p. 1860.
[10] Tarr 1909*b*, p. 2

In some areas the drift is so thin that it fails to mask the irregularities of the underlying bedrock. But in others it has its own distinctive topographic form. End or terminal moraines are distinct from ground moraine on the one hand and from outwash and glacial-lake-floor deposits on the other. Yet, although ground moraine and end moraines ordinarily consist largely of till, some of them are built partly of stratified drift, while some masses of outwash include some till near their upstream ends. These various forms are discussed in detail hereafter. We are concerned now only to point out that there is no simple and general relationship between composition and surface form — and, for that matter, origin.

THE DRIFT IN TERMS OF LAND FORMS

To some extent the drift can be classified according to the various topographic forms in which it occurs. Moraines — end moraines, lateral moraines, and ground moraines, for example — are distinctive topographic features, yet they may include till and stratified drift in varying proportions. Drumlins, though topographically unique, and though consisting mainly of till, may have cores of bedrock and included masses of stratified drift. Eskers, kame terraces, and outwash masses are distinctive forms consisting primarily of stratified drift; yet all may contain some till. Accordingly topographic form as well as composition must be considered in any attempt to describe the drift.

TILL

Composition, Texture, and Structure

Till is a sediment, and it is perhaps more variable than any sediment known by a single name. Its outstanding characteristic is that it is not sorted. It may consist of 99 per cent clay, or 99 per cent large boulders, or any combination of these and intermediate sizes. As is shown below, till takes on the complexion of the near-by bedrocks because it is largely derived from them, and in consequence it varies greatly from place to place. In shale and limestone districts, for example, the till is rich in clay and contains relatively few stones. In districts underlain by granite and quartzite, on the other hand, the till is very stony and the matrix that holds the stones together is poor in clay.

A till rich in clay and with few stones is very resistant to glacial erosion, partly because it is free of joints, so that quarrying is extremely difficult, and partly because it is sticky when moist and thus causes rock particles in the base of the overriding glacier to lodge in it rather than to erode it. Till that has been accumulated by lodgment beneath a

thick glacier is likely to be very tough, having been compacted under a great weight of ice. This toughness is enhanced by the presence of clay which acts as a binder between the fragments of larger size. Tough clay-rich till is often called *hardpan* by well drillers, engineers, farmers, and others who have to cope with it.

R. F. Flint

FIG. 24. Weakly fissile till exposed at Fairhaven East, Connecticut. This till is fine grained, consisting of an abundant matrix of silt, clay, and sand inclosing a few small stones. The vertical channels were made by a power shovel.

Again, a till rich in clay has very little permeability to percolating water. As a result it is rather easily channeled by surface runoff and in extreme cases forms badlands. In contrast, sandy and stony till may be so highly permeable that it absorbs all the rain that falls upon it and thus easily resists erosion.

Although most tills are structureless, some tills, particularly those rich in silt and clay, have a roughly horizontal fissility (Fig. 24) which at first sight looks like stratification but which has the discontinuous flaky quality of light pastry. It is not stratification but structure, and it is probably flow structure induced in the plastic clay under great pressure

as the till was built from the bottom up by accretion from the base of thick glacier ice that flowed slowly over it. This fissility was once thought to be lamination caused by seeping water at the base of the glacier.[11]

FABRIC[12]

We may deduce with T. C. Chamberlin[13] three principal processes in the deposition of till: dumping, pushing, and lodgment. Dumping, a heaping-up process of free sliding and falling, occurs around the margins or on the upper surface of the glacier. Much of the till in ablation moraine accumulates in this way, and it is likely to be coarse grained, because during the dumping process there is usually opportunity for some of the finer grain sizes to be washed away by meltwater.

Pushing is a heaping-up process caused by the snowplow-like action of the rigid terminal margin of a glacier as it advances upon existing coarse deposits. This process can not be very important quantitatively because the weakness of flowing ice is such that the ice would fail after it had pushed up a heap of drift only a few yards in height.

Lodgment is a process of accumulation by accretion or plastering-on. It takes place through the gradual transfer of drift from the basal ice to the ground beneath the ice and is especially effective where the basal load includes a great deal of sticky clay. Stones are pressed into the clay and stick there while the ice flows slowly over them. Most till has accumulated subglacially by lodgment.

The arrangement of the component rock fragments — the *fabric* — in till accumulated by lodgment is not confused, pellmell, and chaotic as it is sometimes said to be. It is organized. The organization is rarely apparent to the eye but is evident only after measurement of the positions of many stones. A very large number of stones lie with their longest dimension parallel with the direction of flow of the glacier that deposited them. A smaller number lie transverse to this direction. Holmes[14] showed that the stones in the former group probably slid along the subglacial floor and then lodged in the upgrowing till, whereas those in the latter group rolled along the subglacial floor just before lodging in the till.

Although the fact that some till has an organized fabric was noticed as early as 1884, till fabrics have been little studied, probably because the study demands arduous digging, classification, and statistical effort

[11] Upham 1891.
[12] See Holmes 1941 and references therein contained.
[13] T. C. Chamberlin 1894, p. 525.
[14] Holmes 1941, p. 1331.

and is not much fun. However, fabrics are important because from them the direction of movement of the ice at the time of till deposition can be inferred with greater accuracy than from striations, drumlins, and other features. Even more important, it can be inferred where more obvious features are absent.

Another concept, widely held, is that the fabric of some tills is the little-altered fabric of the basal drift in the former glacier ice. According to this view the basal ice becomes so fully charged with rock fragments that it loses its plasticity and can no longer flow. The ice overlying it may or may not continue to flow, overriding the stagnant drift-laden ice underneath. Ultimately wastage of the basal ice leaves its drift content essentially with the arrangement it had when incorporated in the glacier, an arrangement that might be called a "fossil glacier fabric."[15]

Shapes of Stones[16]

The shapes of the stones in till are infinitely varied. The vast majority of stones, especially in the harder rock types, have shapes inherited from the shapes of the pieces of rock torn out or picked up by the glacier ice. A great many are bounded by joint and stratification surfaces. Many are raggedly angular, indicating that their sizes are reduced by crushing rather than by abrasion. Many others are rounded, having been worn by streams of glacial meltwater before being picked up by the ice and incorporated in the till, or during ice melting. The prevalence of inherited form accords with common observation of existing glaciers: rock fragments travel for miles on or in a glacier without the slightest modification in their shapes. Stones can be modified only by crushing or by grinding, and these processes can not operate on a stone unless it comes into contact, in flowing ice (not rigid surface ice), with other rock fragments or with the bedrock past which the ice is moving. Therefore stones are altered most readily at the base of the glacier.

Some stones in till have strong radial fractures evidently not inherited from the bedrock but made after the stone had been shaped. These features can result only from crushing, probably between large stones, beneath the weight of a thick glacier.

As might be expected, crushing of rock fragments in transit is far more common in flowing glaciers than in streams of water. In streams attrition is much more common than crushing.

The progressive abrasion of small pebbles of a soft kind of rock

[15] Cf. Slater 1927 and the other listed papers by the same author. The idea is carried to an extreme by R. G. Carruthers (1939).

[16] See Wentworth 1936; Holmes 1944*b*.

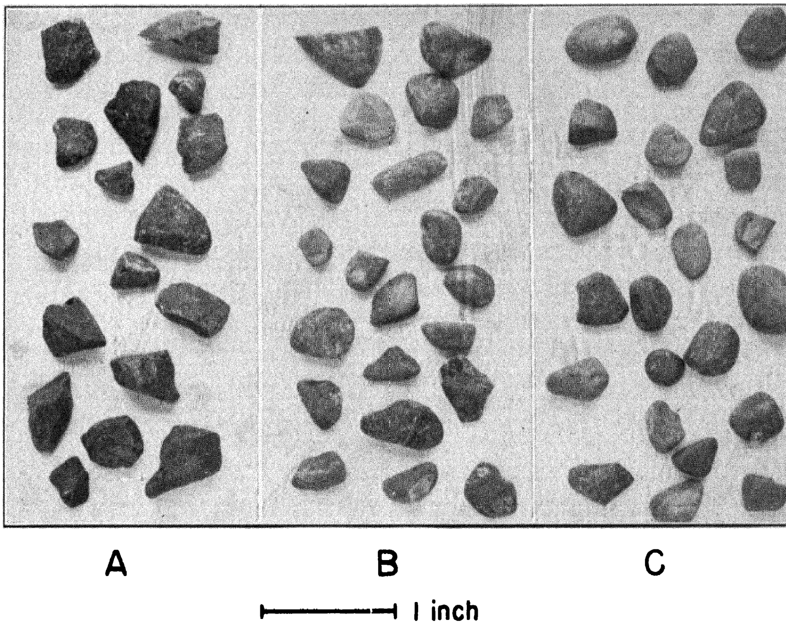


FIG. 25. Progressive abrasion of pebbles during transport in the base of an ice sheet (Holmes).

Small limestone pebbles from till exposed in the region south of Syracuse, New York, at varying distances downstream from the outcrop area of the limestone.

A. About 3 miles downstream.

B. About 8 miles downstream.

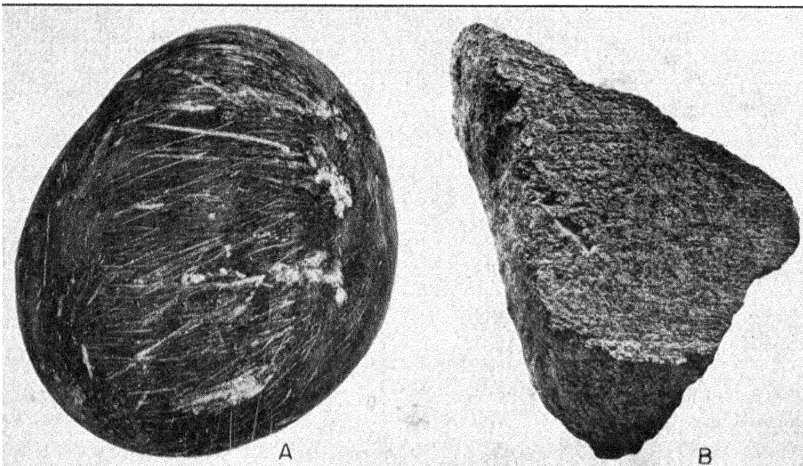
C. About 12 miles downstream.

through a minimum distance of 12 miles of transport is illustrated in Fig. 25.

In most tills a small proportion of the stones are faceted. Their surfaces were ground into facets as the stones were carried forward in the sole of the glacier in contact with hard bedrock beneath. If a stone had ice all around it except at its contact with the bedrock, pressure melting would soon cause it to retract upward into the ice so that it would lose contact with the rock and be worn no further. Therefore it seems likely that during faceting a stone must have been prevented from retracting by the presence of other stones immediately above and behind it.

Some of the facets, especially in soft rocks like limestones, are very flat. Some, especially in harder rock types, are curved like the sole of a well-worn old shoe. The edges between adjacent facets are commonly

rounded. Striations may be present on any part of the surface of a stone. As a rule they have no consistent orientation but run in various directions, crisscrossing each other (Fig. 26*A*). Occasonally, however, stones are found which have a single well-developed facet marked with parallel striations (Fig. 26*B*). The chances are much against this kind of stone having acquired its facet and striations while being carried in the glacier. Probably they were part of the ground over which the ice moved.



C. D. Holmes

FIG. 26. Glacial facets and striations on stones (Holmes 1944*b*).

A. Stone with random striations made while the stone was in transport.

B. Stone with flat facet having parallel striations, both made while the stone was part of a pavement.

Near Toronto, Ontario, an exposure of clay-rich till shows a thin flat-lying zone of stones firmly held in the clay matrix. The stones have beveled upper surfaces forming a common plane.[17] These stones mark an unconformity between two similar sheets of till. The lower sheet was exposed to erosion, and the matrix sediment was washed away, leaving a pavement of stones at its surface. Then renewed glaciation beveled the stones and striated them while the clay matrix firmly held their lower parts. Till accumulated upon the pavement. Features of this kind were first recognized in 1852 by Hugh Miller, who called them *boulder pavements*,[18] not a very good term because most of the stones concerned are smaller than boulders in size.

[17] Coleman 1933, p. 30.
[18] Holmes 1944*b*.

It has been held[19] that the "finished" stone carried and shaped in the base of a glacier is shaped like a flatiron and that all "unfinished" stones tend toward this shape. However, it is very doubtful that the matter is so simple, and the suggestion has not met with general favor.[20].

Till-Like Deposits

Absence of stratification and of size sorting, the two most obvious characteristics of till, are by no means confined to till but are shared with a number of other deposits with which till is sometimes easily confused. Sediments resulting from soilcreep, earthflow, mudflow, and landsliding lack stratification and size sorting. This is true also of "warp," a sediment developed *in situ* mainly by seasonal frost heaving, and of any surficial material reworked by the growth of roots and by the uprooting of falling trees. Unless such sediments involve glacial material they are not likely to include striated or faceted stones or stones from far-distant sources. However, discrimination is sometimes difficult.

Basal Till and Superglacial Till

As long ago as 1877 there were recognized, both in southern Sweden and in eastern North America, exposures of till that seemed to consist of two distinct members. The lower member was dense and clay-rich, with a good many glacially worn stones. The upper member included less clay and silt in proportion to sand, and the stones it contained were on the average larger and less worn than those in the lower member. Torell attributed the lower member to deposition, by lodgment beneath the ice, of the basal drift, and the upper member to the gradual letting-down of superglacial ablation moraine (described in Chapter 5) by slow wastage of the ice beneath it.[21] At that early date he correctly recognized two very different processes of till deposition and thereby established a principle later adhered to widely.[22]

Superglacial till overlying basal till has been identified at an increasing number of localities. The basal till is clay-rich because little or none of the finest sediment it contains had opportunity during deposition to be flushed away by running water. It is dense, tough, and compact because it has been pressed down by the weight of a great thickness of ice.[23]

[19] von Engeln 1930.

[20] Cf. Holmes 1941.

[21] Torell 1877.

[22] Stone 1880, pp. 433–434; Upham 1891; 1895*a*, p. 343; Salisbury 1902, pp. 40–43; E. W. Shaw 1912, p. 8; von Engeln 1929, p. 470; W. B. Wright 1937, p. 26.

[23] The compaction this till has undergone may be inferred from its extreme toughness and may also be deduced from the fact that the pressure on it, caused by the weight of the overlying ice, amounted to about 30 tons per square foot per 1000 feet of thickness of ice.

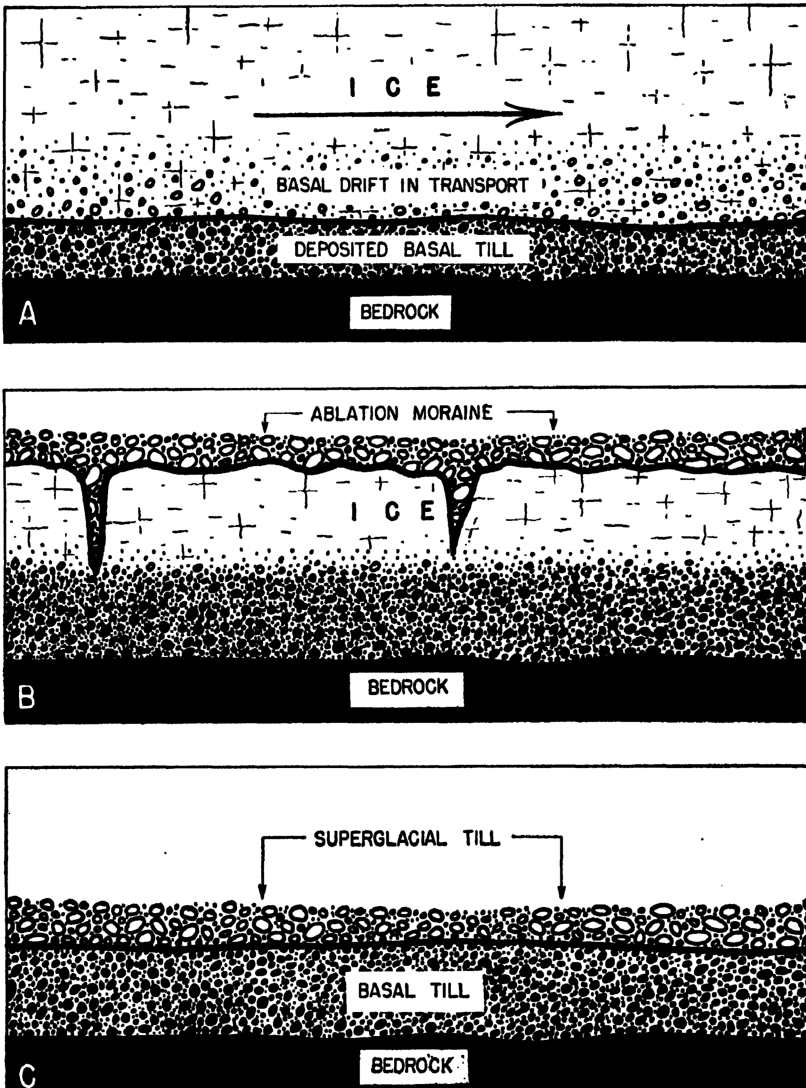


FIG. 27. Comparative origin of basal till and superglacial till.

A. Basal drift in transport lodging over bedrock to form basal till.

B. Later stage; thin ice in the marginal zone is wasting away beneath a cover of ablation moraine. Streams are removing the finest sediments from the superglacial drift.

C. Postglacial condition. The former ablation moraine let down minus some of its finest constituents now forms a layer of superglacial till over the basal till.

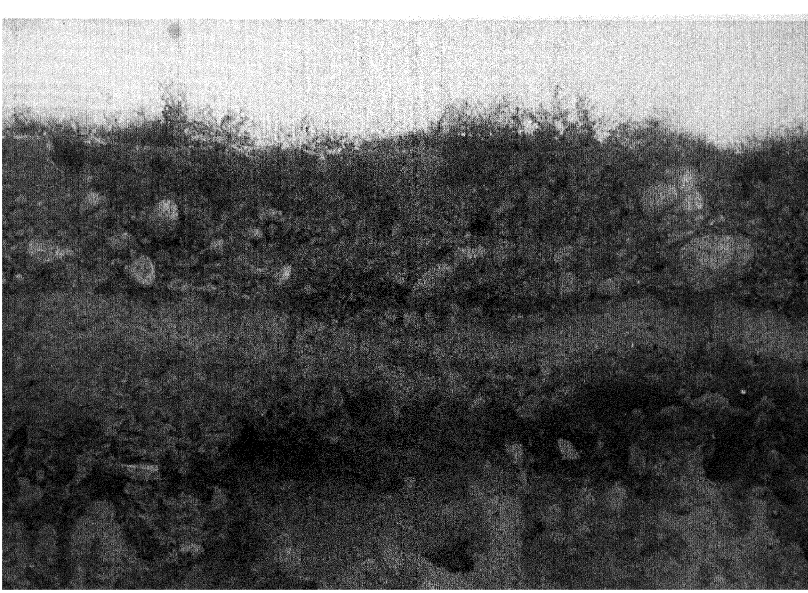
Some of the stones show glacial modification because they were carried in the drift-rich basal part of the glacier where attrition was maximum. The superglacial till, on the other hand, has been subjected to no pressure and little attrition but was repeatedly washed by trickles and rills of meltwater during its existence as ablation moraine. This explains its coarser average grain size, less-worn constituents, and looser texture (Fig. 27). In some places a thin layer of stratified drift separates the two layers of till. Probably this was deposited by meltwater flowing beneath the thin ice in the terminal zone while ablation moraine was accumulating above it.

In Fig. 28 the contact between the two tills appears as a distinct surface; yet it is not an erosional contact. This distinctness is characteristic of the surfaces separating superglacial from basal tills.

An unusual superglacial till is present in many places on Cape Cod and on islands adjacent to it. Many of the stones in it are ventifacts—stones faceted and polished by wind-driven sand and silt. Some of them have not been altered by any other process since the wind cutting came to an end, though others have been rounded as though by stream transport. These facts are best explained by the hypothesis that the stones

FIG. 28. Superglacial till 6 feet thick overlying basal till more than 12 feet thick. Willimantic, Connecticut.

R. F. Flint



were wind-cut while they were part of an ablation moraine, exposed to the full sweep of the wind on the continental shelf. As the thin glacier ice beneath the moraine melted, the superglacial drift was let down upon the basal till without further movement other than slumping, sliding, and reworking by superglacial meltwater streams.[24]

Other combinations exist which resemble the superglacial till-basal till pair. One is the combination of two distinct basal tills of different ages.[25] Another is a single basal till the upper part of which has been decomposed by weathering so that it has a less compact texture than the undecomposed till beneath.[26]

Lithologic Relation to the Bedrocks

The composition of till, stones and matrix together, strongly reflects the character of the bedrocks exposed immediately upstream. Most till, and much of the stratified drift as well, is so closely related to the near-by bedrocks that it is plain that the average distance traveled by a rock fragment from the time it is picked up by the glacier until it is deposited is only a few miles.

W. O. Crosby[27] identified several tons of stones from a small esker near Weymouth, Massachusetts. By comparing them with the bedrock lithology in the region between Weymouth and the Massachusetts-New Hampshire boundary, and computing the results, he found that more than 50 per cent of the stones had been derived from within 10 miles upstream and that 90 per cent of them had been derived from within 20 miles. A similar examination of 300 pebble-size stones from a crevasse filling near Orrington, Maine, revealed that 75 per cent of them were derived from the bedrock in the immediate vicinity.[28]

Eskola[29] identified 961 stones from till in the district south of Riga in Latvia and compared the rock types with the outcrop areas of those same rock types in Finland, the origin of the stones. He made the striking discovery that the percentage of each rock type in his collection of 961 stones was closely comparable with the percentage of the area of Finland underlain by that rock type. Here is his table:

[24] L. R. Thiesmeyer, *unpublished.*
[25] Currier 1941.
[26] Lawrence Goldthwait 1941.
[27] W. O. Crosby 1896, p. 142.
[28] Trefethen and Harris 1940, p. 411.
[29] Eskola 1933.

Rock Type	Percentage Present in Latvian Till Sample	Percentage of Finland's Area in Which This Rock Type is at the Surface
Granitic rocks	64.4	56.5
Migmatites	19.3	21.8
Schists	5.1	9.1
Quartzites and sandstones	2.5	4.3
Calcareous rocks	0.1	0.1
Basic rocks	8.6	8.2
Total	100.0	100.0

These figures do not imply that the Latvian stones were concentrated from the whole area of Finland; actually they could have come only from the southern part of Finland. All the rock types are there, and the Scandinavian Ice Sheet quarried stones from all of them. The stones came from distances of at least 200 miles, whereas much of the till matrix was derived from the much softer sedimentary rocks of Estonia, Latvia, and the Baltic Sea floor. At first this long-distance travel seems to contradict the statement that the till is of local origin. But it does not, for here only stones are involved; the matrix, actually far more abundant than the stones, is not taken into account. Furthermore the stones are of notably durable types. The table suggests that the weakest among them, the schists and the sandstones, clearly suffered most in transit.

As a matter of fact the far-traveled stones in till are always of durable types, largely because of the prevalence of glacial crushing, already mentioned. A process of natural selection takes place in transit, resulting in a "survival of the toughest." It is not uncommon to find tough stones and boulders as much as 500 or 600 miles from their sources, and a few have traveled even farther[30] (Fig. 30).

Not only does till reflect the local bedrocks; it reflects them in a definite way. Conspicuously jointed hard rocks such as granites yield tills consisting of boulders and large stones in a scanty matrix of pronounced sandiness. Sandstones yield stones in an abundant matrix rich in sand. Limestones and dolomites yield boulders and smaller stones (torn out by quarrying) in an abundant matrix of clay and silt consisting in large part of minute fragments of limestone. Shales yield tills that are largely clay. The stones present are mainly rocks other than shale and are derived from elsewhere.

That two tills at the same locality, only slightly different in age and both deposited by ice moving in the same direction can be very different

[30] Cf. Antevs 1928*a*, pp. 65–67.

in composition is shown in the Cape Cod district.[31] The Montauk till there has an abundant matrix rich in silt and clay. The depositing ice passed across a belt of the interglacial Gardiners clay and incorporated much of this clay into the Montauk till. On the other hand the slightly younger "later Wisconsin" till is very stony and has a scanty matrix consisting of sand with very little silt or clay. Between the two glaciations the Gardiners clay in this district had become covered with a sand formation. The later glacier, accordingly, picked up sand but very little clay. Such differences, once established within a given district, are very helpful in identifying and mapping a succession of tills.

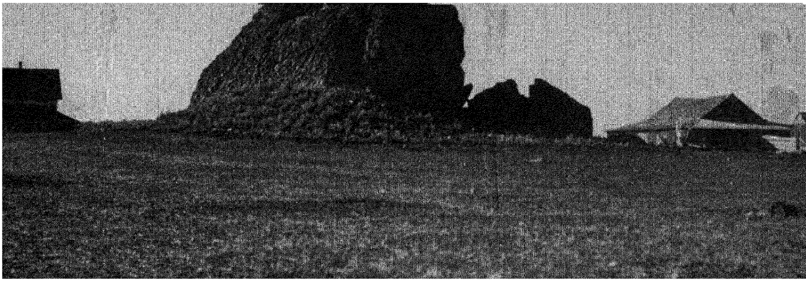
Since sheet structure in granites becomes more widely spaced with increasing depth, the size of glacially quarried monoliths increases with the depth below the preglacial surface from which they were derived.[32] Hence a till in which the granitic stones are of pebble and cobble sizes suggests erosion of a superficial zone, whereas large boulders suggest deep erosion. Hence a till including only moderate-size stones overlain by a deposit of large granitic boulders, a relationship that occurs in the Long Island-Cape Cod region, may mean an earlier glaciation that removed the superficial material in the region to the north, followed by a later glaciation in which only the formerly deep-seated rock in the lee slopes of hills was available for erosion, the mantle having been already removed and the intervening time having been too short for the development of another mantle by weathering.

ERRATIC STONES AND BOULDERS

Every till contains some stones that are derived from bedrock different from the rock on which the till rests, and, where till is scanty or absent, stones and boulders of this "different" character are commonly found lying free on the surface. Whether free or embodied in till, these "different" rock fragments have been called *erratics* ever since Charpentier (1841) described as *terrain erratique* the stones and boulders of Alpine origin lying scattered on the southern slopes of the Jura Mountains.[33]

Many erratics are very large (Fig. 29). The Madison erratic near Conway, New Hampshire, is a chunk of granite measuring 90 by 40 by 38 feet and probably weighing 10,000 tons. The Ototoks erratic 30 miles south of Calgary, Alberta, is a quartzite derived from at least 50 miles to the west. It consists of two pieces which together measure 160 by 55 by 25 feet and have a calculated weight of 18,150 tons. More remarkable

[31] Woodworth and Wigglesworth 1934, pp. 70–71.
[32] Jahns 1943.
[33] According to North (1943, p. 5) erratic blocks were first mentioned by De la Beche in 1819.



U. S. Bureau of Reclamation

FIG. 29. Erratic boulder of basalt on ground moraine. Columbia Plateau west of Coulee Dam, Washington.

although slightly smaller is the great tabular erratic of Silurian limestone resting on Illinoian till in Warren County, Ohio. It has an area of more than 20,000 square feet; yet its thickness averages only 5 feet. Its weight approximates 13,500 tons, and it was ice-transported at least 4.5 miles. How this was accomplished without breaking the monolith has never been explained.

Although a good many erratics have traveled hundreds of miles, in both North America and Europe, no erratics traced to their apparent sources in the bedrocks are known to have traveled more than 700 miles.[34] None of those in central and eastern United States have been shown to have been derived from north of the southern fringe of the Canadian Shield (Fig. 30).

INDICATORS

If a stone is foreign to the local bedrock it is an erratic. If, in addition, its place of origin is known by direct comparison with the bedrock there, it is an indicator as well. By this term[35] "is meant the stones . . . the characteristics of which are so peculiar and distinct that it can be determined exactly from what spot or rather limited area of the region once covered by the ice they came." Indicators were first recognized by Saussure, a sharp-eyed observer, whose classic description of them[36] has never been improved upon.

In general a line drawn from the locality of an indicator to the locality of its origin parallels the striations and other evidences of direction of glacial flow between the two points. On the other hand the path of travel does not necessarily approximate a straight line inasmuch as the ice may have changed its direction of flow between the time the indicator was picked up and the time it was deposited in its present locality.

[34] Antevs 1928*a*, p. 65.
[35] Milthers 1909, p. 1.
[36] Saussure 1786–1796, vol. 1, 1787, p. 201.

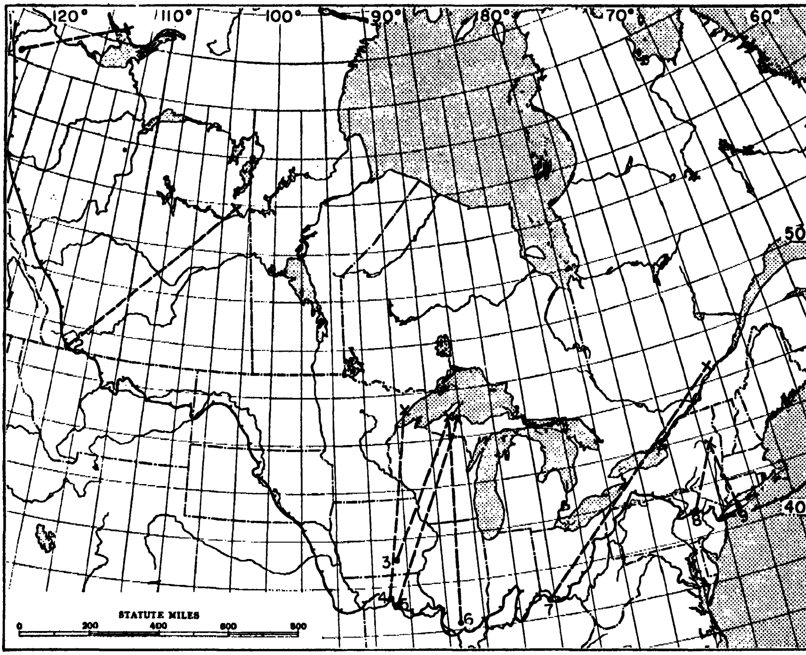


FIG. 30. Apparent paths of selected far-traveled erratics in North America. From supposed source localities (×) to points of occurrence (●). Rock types: (1) cordierite-biotite schist, (2) "Pre-Cambrian," (3) native copper, (4) anorthosite, (5) jasper conglomerate, (6) native copper, (7) norite gneiss, (8), (9) granite gneiss. Hatched line = approximate limit reached by ice sheets. (Data on erratics from Antevs 1928*a*, and other sources. Base map courtesy American Geographical Society.)

After the numbers of indicators from various known sources, in tills exposed at many localities, have been counted, statistical methods have been found useful in discriminating between successive till sheets that are otherwise much alike. In any one district slight differences in the direction of flow of successive ice sheets can thereby be detected, provided the sources of stones are known with sufficient accuracy. Stone counts therefore, under suitable conditions, have a distinct stratigraphic value.[37]

PERCHED BOULDERS

Many large erratics (and also boulders that are not erratics) lie on the tops of hills or bosses in such unstable positions that they are easily dislodged. Such boulders are said to be *perched*. Most of them owe their positions to accident. At every point on the ground the motion of a wasting and thinning ice sheet decreases gradually until sooner or later

[37] See Milthers 1909 (a basic paper); 1936; also a critical summary in Woldstedt 1935*a*.

it reaches zero. At some points this is bound to occur at a time when a boulder is perched there. The ice melts and frees it in the perched position. A great many more almost-perched boulders, with even greater instability, slid or rolled away when released by melting and reached stable positions. These attract no special notice, but they are vastly more numerous than the boulders still in perched positions.

Boulder Trains

Detailed study is bringing to light more and more *boulder trains*—conspicuous linear strings or fan-shaped spreads of stones and boulders of a single rock type glacially moved from their sources in the bedrock. Some are free on the surface; others are incorporated in the till. They range from a few hundred yards to hundreds of miles in diameter. Their importance lies in the fact that they are perhaps the best available index of the local direction of flow of an ice sheet.

The simplest boulder trains consist of a line of boulders lying on the surface and oriented in the direction of flow. A conspicuous example is the Snake Butte train in north-central Montana.[38] A more commonly reported type consists of a group of stones and boulders lying, chiefly in the till, within a fan-shaped area whose apex is the locality of origin of the pieces.[39] More boulder trains have been identified in New England than in central North America because bedrock types are diverse and the region is but thinly covered with drift. Figure 31 shows the principal trains. Nearly all are fan-shaped, and some are very extensive.[40]

Probably all boulder trains begin by having a linear form. The fan shape results from change in the direction of glacier flow, induced either by the increased influence exerted by topography as the ice sheet thinned or by shift in the accumulation of snowfall during the growth and decay of the ice sheet. The width of arc of the fan is determined by the arc through which the shift in direction occurred after erosion of the rock at the source of the fan began. Probably most linear trains were made at so late a date that no shift in direction of flow occurred subsequently.

One of the largest and most famous trains is the Ailsa Craig train[41] (Fig. 32) consisting of riebeckite-eurite, a very distinctive fine-grained granitic rock found uniquely in Ailsa Craig, an 1100-foot bun-shaped island in the Firth of Clyde, off southwestern Scotland. Quarried from the island in great quantity by Scottish glacier ice, these rock fragments

[38] Knechtel 1942.

[39] For representative descriptions see Shaler 1893; Hausen 1912.

[40] For boulder trains in the Great Lakes region see Martin 1932, pp. 253, 402; Hobbs 1899. For Europe see Milthers 1909; Sauramo 1929, fig. 14.

[41] W. B. Wright 1937, pp. 67–68.

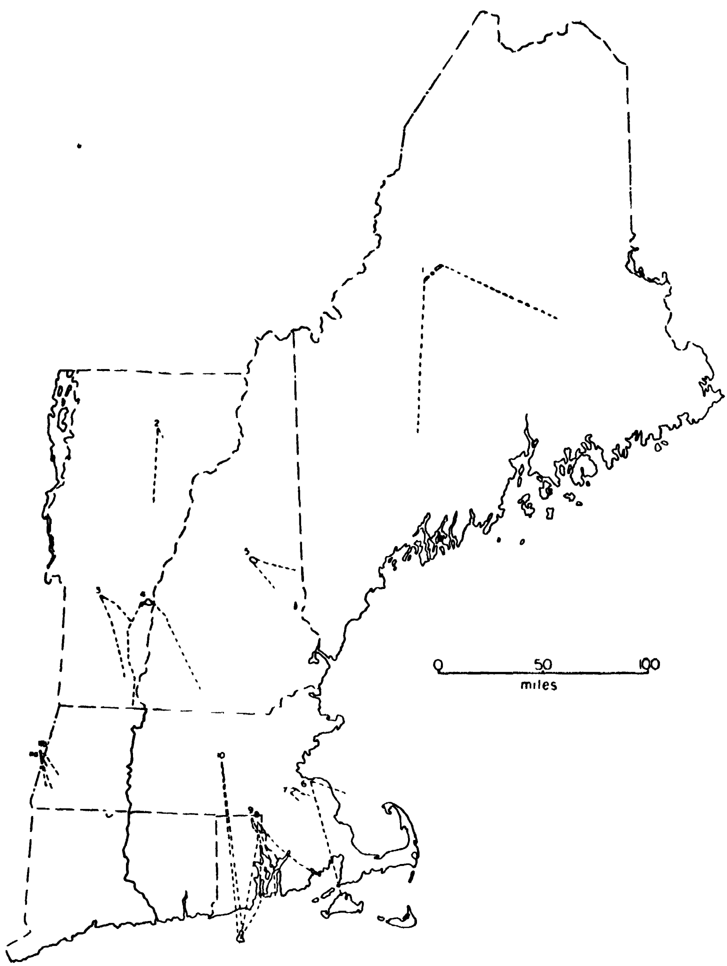


FIG. 31. Fan-shaped boulder trains in New England. (Compiled from various sources including J. W. Goldthwait *in* Flint and others 1945.)

1. Mt. Kineo (felsite).
2. Burlington (quartzite).
3. Cuttingsville (syenite).
4. Mt. Ascutney (quartz syenite).
5. Red Hill (syenite).
6. Hingham (red felsite).
7. Randolph (sandstone and shale).
8. Diamond Hill (agate).
9. Iron Hill (peridotite).
10. Georges Hill (chiastolite).

11*a*, *b*. Richmond-Great Barrington (amphibolite schist).

were carried down the Irish Sea floor area. They were deposited along the east coast of Ireland as far south as Cork and along the Welsh coast opposite. The train is nearly 300 miles in length and some of its stones

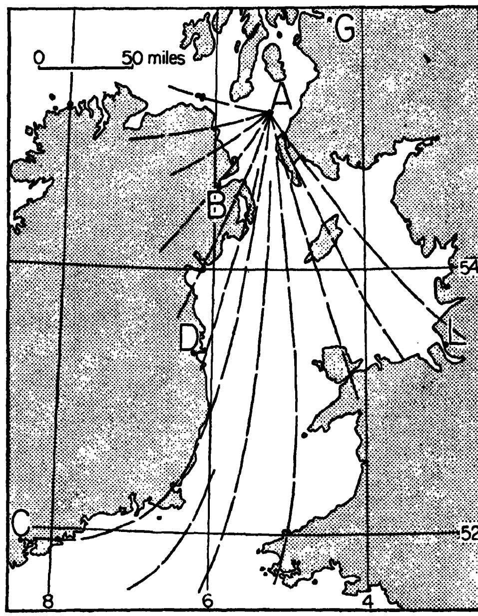


FIG. 32. Ailsa Craig boulder train. The wide angle of the fan indicates unusually great shifts in direction of flow during glaciation. *A*, Ailsa Craig; *B*, Belfast; *C*, Cork; *D*, Dublin; *G*, Glasgow; *L*, Liverpool.

have been carried upward at least 250 feet vertically above the highest point of Ailsa Craig. Only the wings of the fan are visible. The great bulk of it is submerged beneath the Irish Sea.

DRUMLINS[42]

FORM, COMPOSITION, AND DISTRIBUTION

It is hardly possible to define a drumlin more exactly than to say that it is a streamlined hill of glacial drift. The word *drumlin* is Irish[43] and was first used as a general term for hills of this kind by Maxwell Close in 1866.[44] It is now common to all languages. The ideal drumlin form is half-ellipsoidal like the inverted bowl of a spoon, with the long axis

[42] Good general references are Hollingworth 1931; Ebers 1926; 1937; Fairchild 1907; 1929.

[43] Gaelic *druim*, the ridge of a hill.

[44] Close 1867.

paralleling the direction of flow of the former glacier. The steeper, blunter end is the stoss end, and the tapering end is the lee end. Many drumlins, however, vary from the ideal, grading all the way to rounded drift hills of no systematic form whatever.

The ideal drumlin is nearly a mile long, 1200 to 1800 feet wide, and 60 to 100 feet high. But all variations from less than half a mile up to several miles in length and from less than 20 feet up to more than 200 feet

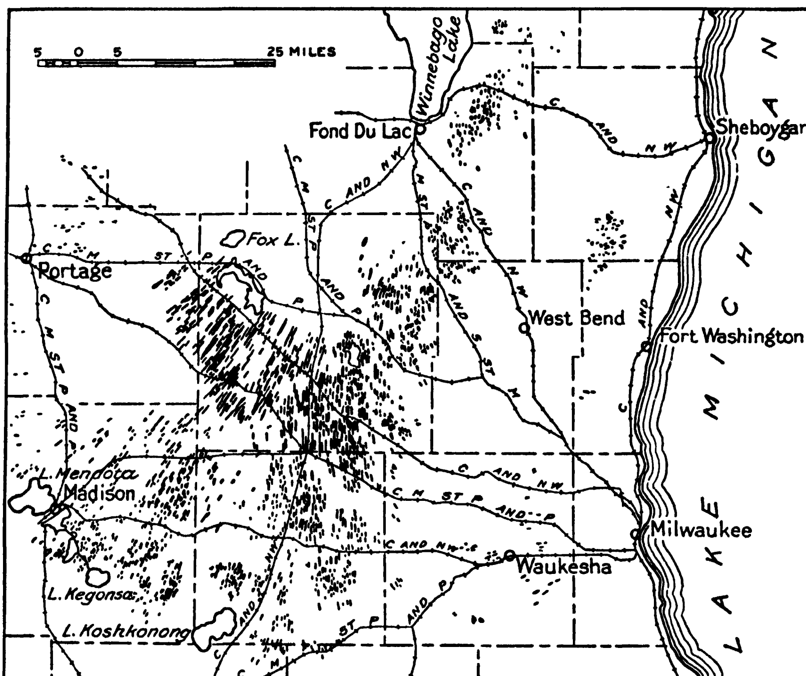


FIG. 33. Drumlins molded by the Green Bay glacier lobe in southeastern Wisconsin (Alden). Radial flow of former ice is recorded by the orientations of the drumlins.

in height are known. Proportions range from nearly circular to very long and narrow. Single isolated drumlins are almost unknown; usually drumlins occur in great "fields" running into the hundreds. In some places adjacent drumlins are indistinctly separated from each other. They form "double or triple ridges united at the steeper end with the tails only distinct; doublets en echelon, the tail of one rising from the flank of the other as an inclined terrace or shelf; small drumlins plastered on the side of larger ones, giving a grooving effect to the flanks of the latter; two-tiered drumlins . . .".[45] Despite these variations, however, the

[45] Hollingworth 1931, p. 325.

long axes parallel the former ice flow. This is strikingly evident where radial flow occurred in the terminal part of a glacial lobe (Fig. 33).

The majority of drumlins are composed of clay-rich till similar to the till in the areas between the drumlins and packed hard so that it is extremely tough. Some, however, consist of sandy till. Most exposed sections show the till to be structureless, though a faint structure parallel with the surface has been observed in some clay till. Regardless of the kind of till, masses of sorted and even stratified sediment occur in it, and a few drumlins appear to consist largely of sorted drift. Although many drumlins consist of drift only, some have cores of bedrock, usually

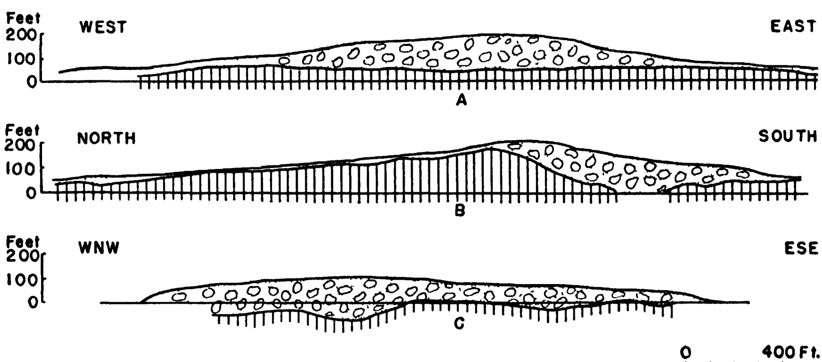


FIG. 34. Sections through three drumlins showing the bedrock surface beneath the till. Horizontal and vertical scales are identical. The irregularities in the surfaces of the drumlins are typical.

A. Oblique section through Mt. Ida, Newton, Massachusetts. The drumlin trends northwest-southeast. Bedrock surface reconstructed from borings and a tunnel. (Redrawn from I. B. Crosby.)

B. Transverse section through Parker Hill, Roxbury, Massachusetts. Bedrock surface reconstructed from borings and a tunnel. (Redrawn from I. B. Crosby.)

C. Long section through Governors Island, Boston Harbor, Massachusetts. Steepening of the surface profile at the two ends is the result of erosion by waves. Bedrock surface reconstructed from seismic soundings. (Section constructed from data published by F. W. Lee.)

though not universally at or near their stoss ends (Fig. 34). These are streamlined variants of the crag-and-tail hills mentioned earlier. And in some areas the drumlins consist of bedrock thinly veneered with till. These have been referred to as *rock drumlins* and regarded as distinct from true drumlins, although on a basis of form alone the two types are indistinguishable.

From all this it is plain that in drumlins there is complete gradation from rock to drift, and that this gradation is independent of outer form. Most drumlins, whether consisting of drift or of rock, occur in districts

where the bedrocks are rich in clay or in clay-forming minerals. The presence of all types in a single field suggests that all were made contemporaneously under an all-embracing set of conditions.

Drumlins are widely distributed in both North America and Europe, and, though conspicuous forms constitute rather distinct "fields," less ideally shaped drumlins are present in intervening districts where they have escaped attention. The most conspicuous drumlin fields in North America lie in central-western New York (about 10,000 drumlins), east-central Wisconsin (about 5000 drumlins) south-central New England (about 3000 drumlins many of which are rock drumlins), and southwestern Nova Scotia (2300 drumlins). Large groups of remarkably long, narrow forms with pointed stoss ends and consisting of very sandy till occur in northern Saskatchewan and Manitoba. These are referred to in the early literature by the local Indian name *ispatinow*. An irregular field of long narrow forms very thickly coated with loess occurs in eastern Iowa and northwestern Illinois. In the early literature these, too, were called by a regional Indian name, *paha*.

Because some drumlin groups lie a few miles or tens of miles back (upstream) from end moraines that mark the terminal positions of glacial lobes[46] it has been inferred that this is a systematic relationship implying comparatively thin ice at the time and place of drumlin making. But this is not a definite requirement for drumlin production, as in at least one "field" the long axes are not at right angles to the successive termini of the shrinking glacier.[47]

Origin

The forms of both true drumlins and rock drumlins are clearly streamlined, offering a minimum of resistance to the ice flowing over and past them. This indicates that they were shaped by the flowing ice, an inference on which opinion is virtually unanimous. The shaping, however, could have taken place either through the erosion of pre-existing hills or through the progressive lodgment of drift around some nucleus. It is obvious that rock drumlins must be mainly the result of erosion, but the consensus is that in most true drumlins accretion has been the predominant process. Either way the streamline form would result.

It is believed that true drumlins are made beneath actively flowing, expanding glaciers rather than during shrinkage. This belief is supported in part by the parallelism of drumlins throughout a wide "field," which would not occur if the ice were thin and waning, for then topography would cause the flow to diverge locally from the main regional direction.

[46] Alden 1918, p. 253.
[47] Hollingworth 1931, p. 341.

The locus of drumlin making is thought to be the submarginal zone, for there the ice is comparatively thick and the basal ice is normally charged with a heavy load of debris. The greater the basal load the less plastic the basal ice. Because friction between clay and clay is greater than that between clay and ice, clay-rich drift tends to lodge on the subglacial surface, building up a sheet of till there. Where knobs and bosses of bedrock project somewhat, basal drift is lodged around them. Elsewhere the mass of basal drift may be so great as to cause the ice that holds it together to stagnate while the less heavily loaded ice around it continues to flow over and past. In this way a nucleus is formed without benefit of bedrock, and drift is lodged by accretion upon it because clay readily sticks to clay. At all times the growing subglacial mound has a streamline form.[48]

It has already been said that some drumlins have cores of bedrock. One point of view carries this idea still further and says that all drumlins have cores, many of the cores being till.[49] The drumlins strongly cliffed by the waves of Lake Ontario provide admirable sections, whose structures seem to bear out the core hypothesis. They characteristically exhibit nuclei of massive till overlain gradationally by much more variable material, with partings and other discontinuities paralleling the drumlin surfaces. These facts are consistent with the general theory of drumlin accumulation suggested above.

The smoother the subglacial surface and the more evenly distributed the drift in the base of the ice, the greater will be the tendency to deposit a uniform sheet of till, free of drumlins and other conspicuous irregularities. But with rock bosses present and with irregular distribution of basal drift, especially if the drift is rich in clay, drumlins will tend to form by progressive accretion. The drumlins built of sandy till have not been studied fully enough to enable us to say how the process differs when the building material includes a large percentage of sand. It is perhaps significant that these sandy drumlins are extreme in form, being very long and very low in proportion to their widths.

The sandy drumlins may be low, but no drumlin is much more than 200 feet high and not many are more than half that height. Very possibly this fact reflects the upper limit of drift in the basal ice. Accretion could occur as long as the ice continued to bring drift into contact with the growing nucleus. But when the drumlin was built up to a height at which no more drift was brought to it, further upward growth would cease. Time available for building may also have been a factor, though it can hardly have been more than secondary.

[48] The elements of this theory of origin were first stated by I. C. Russell (1895).
[49] Slater 1929*b*.

Once formed, drumlins are not easily altered or destroyed by erosion. Tough clay-rich till is always very resistant to glacial erosion because it is nearly free of joints and hence not susceptible to plucking, and because its rounded form reduces abrasion to a very slow process. Hence a regional shift in the direction of glacial flow is reflected only slightly and after some time by changes in the form of drumlins already built.

For this reason the long axes of drumlins are not particularly delicate indicators of the direction of flow of a glacier. In general they parallel the regional directions of striations, of boulder trains, and of eskers.

MORAINE

The word *moraine* is an ancient French word long used by peasants in the French Alps for the ridges and embankments of earth and stones around the margins of the glaciers in that region. It appeared in the literature as early as 1777 and was taken up and used by Saussure and later by Venetz and Charpentier, and was given wide currency by Agassiz. The recognition, later, of a wide variety of forms of drift fashioned by large ice sheets made it necessary to extend the original quite limited meaning of the word. Accordingly we now think of moraine as an accumulation of drift with an initial topographic expression of its own, built within a glaciated region chiefly by the direct action (deposition and thrust deformation) of glacier ice.[50] Moraine is usually subdivided into ground moraine, end moraine, lateral moraine, medial moraine, and ablation moraine.

Ground Moraine

Ground moraine is relatively widely distributed moraine, ordinarily thin compared with its areal extent and usually with gently irregular topographic expression. Rarely more than a few tens of feet in thickness, it forms undulating plains with gently sloping swells, sags, and closed depressions, the whole having a local relief of no more than 20 to 30 feet. The predominant material is till, though stratified drift is present in places. The till is thought to have accumulated largely by lodgment beneath the ice but partly also by being let down from the upper surface of the ice through the ablation process already described.

Ground moraine is the surface material throughout many thousands of square miles south and west of the Great Lakes and in the Great Plains region of west-central Canada, as well as in many parts of northern Germany, Poland, and western Russia. In many places it is

[50] In Scandinavian literature *moraine* is often erroneously used as a synonym for *till*.

broken by hills of bedrock that project through it and, as in New England, by areas of till so thin that it lacks topographic expression other than that of the immediately underlying bedrock surface.

End Moraine[51]

An *end moraine* (also known as a *terminal moraine*) is a ridgelike accumulation of drift built chiefly along the terminal margin of a valley glacier or the margin of an ice sheet. It has a surface form of its own and is the result chiefly of deposition by the ice, or deformation by ice thrust, or both.

The essential thing about end moraine is its close relationship to the terminus or margin of the glacier. Many end moraines are no more than a marginal thickening of the ground moraine.

The ridge that is the typical form of an end moraine is more or less discontinuous. It is broken by gaps, most of which mark the positions of meltwater streams that were usually contemporaneous with the building of the moraine and some of which persisted long afterward. Some of the gaps, however, appear to mark sectors in which no moraine was ever built, either because of lack of debris in the ice or because the terminal zone of the glacier had become stagnant at the time. The ground-plan form of most end moraines is crescentic, recording the lobate margin of the glacier that built them. In some regions the crescents are multiple, giving the moraines the appearance of festoons, clearly shown on many glacial maps. End moraines of valley glaciers characteristically merge in the upstream direction with lateral moraines, forming with them a continuous marginal rampart (Fig. 6).

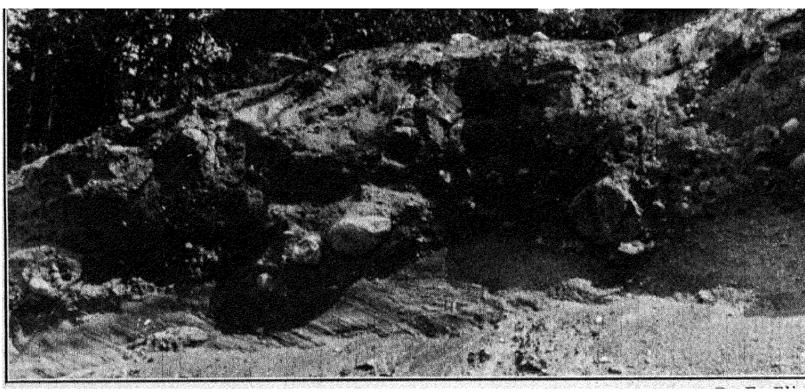
The end moraines of ice sheets, common in central North America and in central Europe, rarely exceed 100 to 150 feet in height, though the two subparallel bases of each moraine may be several miles apart. Some moraines have a single crest; others have several, indicating that they are in reality several successive end moraines built in such close proximity that together they form a unit. The end moraines of many valley glaciers in mountain regions are much higher and steeper, often reaching several hundred feet and rarely even more than 1000 feet in height.

To some extent the greater height and steeper slopes of the end moraines of some large valley glaciers in mountain regions as compared with those of ice sheets probably result from the steep gradients, rapid flow, and rapid erosion associated with valley glaciers. A greater load of drift scoured from the valley floor and avalanched from the valley sides is spread through a thicker zone at the base of the glacier. Thus an initially

[51] See Gripp 1938.

high morainic ridge is likely to result. The ridge is added to rapidly because rapid flow permits the basal ice to carry its load high on to the growing moraine before ablation brings about deposition. Sluicing away of fine sediments is more effective in steep mountain valleys than on lowlands; hence the residual mass is likely to be coarse, permeable, and resistant to erosion.

The clay-rich till of most lowland regions, however, forms lower moraines partly because it is built up chiefly by lodgment beneath the ice terminus, and hence is low to begin with. More significant, its high content of clay makes it yield readily to solifluction, which further reduces the initially gentle slopes.



R. F. Flint

FIG. 35. Detail of northern (stoss) slope of Salpausselkä moraine at Nummenkylä, Finland, showing till overlying delta-bedded sand.

Though most end moraines consist of till, all contain some stratified drift and some consist predominantly of stratified material. Of the latter some, like the famous Salpausselkä ridges (Fig. 35) in southern Finland, are stratified mainly because they were built where the ice sheet ended in ponded water and the drift it delivered at its terminus was promptly sorted and deposited chiefly in the form of a row of deltas. Others, like the ridges that form the cores of the moraines on Long Island and Cape Cod, were evidently formed on land as rows of alluvial fans, and they indicate the rapid production of meltwater capable of handling most of the drift delivered by the ice sheet to its marginal zone.

Regardless of composition the concentration of drift in the narrow belts that are end moraines indicates that the glacier margin occupied these positions through long periods compared with its occupation of neighboring positions. At the times when end moraines were built the

rate of wastage balanced the rate of outward flow of the ice; the part of the glacier concerned was essentially in equilibrium.

There is little direct evidence on the matter, but what little there is suggests that most bulky and extensive end moraines are the result, not of pauses during the general shrinkage of a glacier, but of the culminations of re-expansions during general shrinkage. Also it is probable that an end moraine once partly built is self-extending. That is, the obstacle created by the growing moraine diminishes the average rate of flow of ice in the terminal zone, exposes the ice to ablation for a longer time back of the moraine, and so causes greater accretions of drift to the moraine. The higher and more massive the moraine, the greater the obstacle. Thus overriding of the moraine becomes increasingly difficult, while any slow shrinkage of the glacier that may take place merely adds to the width of the moraine. Only when shrinkage becomes rapid does growth of the moraine cease.

The topography of end moraines in detail varies from a smooth, gently undulatory surface common in moraines built of clay-rich till to a sharply irregular surface marked by knolls, hummocks, and closed depressions, common in moraines in which coarse-grained till and stratified gravel are conspicuous. Many of the gravel knolls are small alluvial fans built out from the front of the ice by little streams of meltwater and then destroyed by collapse of their upstream parts when the ice later melted away.

As a general rule the proportion of stratified drift to till in an end moraine reflects the ratio between rate of flow and rate of melting of the ice. Slowly flowing but rapidly melting ice produces much stratified drift, whereas rapidly flowing but slowly melting ice produces much till. Probably this is one reason why the moraines built near the maxima of the last great ice sheets in North America and Europe are more bulky and include more till than the moraines built later, when the ice sheets had shrunk to a considerable extent.

End moraines are so variable in form and composition that they include very unusual types. Certain low end moraines are believed to have been formed not by deposition of drift but by local ridging-up of a sheet of previously deposited drift by thrusting action of the glacier terminus, and also by presssure resulting from the weight of the overlying ice. Moraines of this kind are essentially structural rather than depositional features.[52] Many moraines in northern Germany, especially, are said to have been formed in this way. They are called *Stauchmoränen* (thrust moraines).

Some end moraines are composite, having been built during one glacial

[52] M. L. Fuller 1914, pp. 201–207; Gripp 1938.

age and added to during a later glaciation. The addition commonly consists of a thin veneer of till which conceals the older drift beneath and creates the false impression that the later glacier was capable of constructing massive moraines. This is seen to have been far from the truth when the unconformity separating the superficial drift from the main mass of the moraine is exposed. Examples are the Fort Wayne, Middle, and Defiance moraines in the Ann Arbor district, Michigan.[53] Others are the Harbor Hill moraine on Long Island[54] and the Cape Cod moraine on Cape Cod,[55] both of which consist of rows of massive outwash fans, veneered with till during later expansions of the ice sheet that completely enveloped the heads of the fans. Still other end moraines are described as built by combinations of thrust and deposition.

Some end moraines have been described as *interlobate,* that is, built along the lines of contact between the lateral margins of two adjacent lobes of an ice sheet. These generally consist mainly of stratified drift.

End moraine does not necessarily occur at the outer margin of a drift sheet. As a matter of fact the outer limit of the youngest or Wisconsin drift sheet in North America from the Atlantic Ocean to the Rocky Mountains is marked by end moraine throughout not much more than half of its total extent. This drift is so recent that the moraines can not have been destroyed by later erosion. They simply never were built. In a few places an end moraine was not built because meltwater was so abundant and flowed so rapidly that the drift was sluiced away as fast as it was delivered at the glacier terminus.[56] This was notably true of glaciers occupying mountain valleys with steep gradients. In most places, however, the fact seems to be that the margin of the glacier did not stay at its farthest line of advance long enough to build an end moraine. Equilibrium in the glacier was reached only momentarily if at all.

In some districts[57] successive small moraines occur in series. Consisting chiefly of till, these moraines are 5 to 20 feet high and some have unbroken lengths exceeding a mile. Their spacing varies, but they are commonly 100 to 300 yards apart. Probably small end moraines of this kind are made chiefly by the push process. They have been regarded as annual, each ridge made in an individual year during a period of general shrinkage of the glacier. Although this has never been proved it is not at all improbable.

[53] Leverett 1908.
[54] Fleming 1935.
[55] Mather, Goldthwait, and Thiesmeyer 1940.
[56] Knopf 1918, pp. 96, 103; Eugenjusz Romer 1929, p. 172.
[57] Southern Finland (Sauramo 1918, p. 27), southern Sweden (De Geer 1940, p. 112, fig. 34); central North America (Norman 1938; Gwynne 1942).

Ablation Moraine

The gradual accumulation of drift on the surface of the thin terminal part of a rapidly wasting glacier was described in Chapter 5. A deposit of drift let down irregularly from the surface of the glacier onto the ground by gradual ablation of the intervening ice is ***ablation moraine.*** This name was first used by Tarr[58] although the origin of this kind of deposit had been recognized in 1879 by Penck, who called it *Oberflächenmoräne*. Its appearance while still resting on glacier ice is one of chaotic ridges, knolls, and depressions[59] though this strong relief is mainly in the ice surface itself and not in the drift. After the drift has been let down onto the ground the relief is reduced, and sometimes almost eliminated, particularly where the deposit consists chiefly of till. Where the material was reworked into stratified drift in the presence of the last wasting remnants of the glacier ice, the surface form is likely to be one of knolls and depressions.

The repeated slumping and sliding of drift on slopes of melting ice affords so much opportunity for washing by meltwater that much ablation moraine consists only of the coarser rock fragments originally present in the drift, the finer elements having been flushed away. In a similar manner stratified drift may be deposited on thin ice and subsequently let down onto the underlying surface.[60] The ice must be very thin in order that the process of collapse does not destroy the stratification.

So complete is the coating of ablation moraine over the marginal zone of the Malaspina piedmont glacier in coastal Alaska that an extensive forest has grown up on it. Only here and there does slumping reveal the presence of glacier ice not far below the surface.[61]

Rock Glaciers[62]

A feature common in some glaciated mountain districts and related to ablation moraine is the *rock glacier* (Fig. 36). In form it resembles a small valley glacier; its distal area is marked by transverse concentric ridges, and it heads in a cirque. It consists of ragged, angular nonsorted rock fragments, many of them very coarse, derived from the walls of the cirque. In some rock glaciers interstitial ice is present as a cement between the rock fragments.

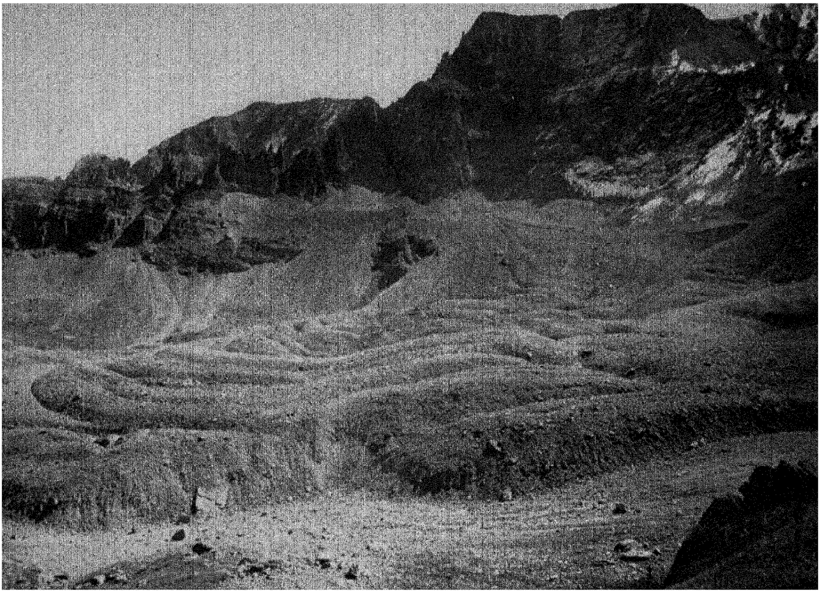
[58] Tarr 1909*a*, pp. 85, 88.

[59] See the descriptions by Tarr 1909*a*, pp. 85–88; Ahlman 1933*a*, p. 173; Mannerfelt 1945, pp. 12–16.

[60] Cf. Gripp 1932, p. 24.

[61] I. C. Russell 1893.

[62] See Capps 1910; summary in Parsons 1939, pp. 745–747. These features have also been called *rock streams*.



Whitman Cross, U. S. Geol. Survey

FIG. 36. Rock glacier at head of Silver Basin near Silverton, Colorado.

Various theories of origin of rock glaciers have been advanced. The one that has met with general favor ascribes the debris to avalanching from the cirque walls onto the surface of a waning true glacier. The resulting thick superglacial drift is slowly carried forward and deposited as successive ridges of end moraine. Even after wastage causes the underlying glacier ice to disappear the formation of interstitial ice from local precipitation may induce continued slow movement in the mass through a long additional time.

According to this theory most rock glaciers consist essentially of end moraine and ablation moraine, which has been subjected in some cases to postglacial movement.

SUMMARY

In summary, we have described the various forms of till and have indicated the more important inferences drawn from till as to the glaciers that deposited it. We have discussed briefly the several types of moraines, pointing out the fact that till is not the only material of which a moraine may be composed. We now turn to a discussion of stratified drift, keeping in mind the fact that the relation of till to stratified drift is completely gradational.

Chapter 8

GLACIAL DRIFT II. STRATIFIED DRIFT

ADVANCE SUMMARY

Accumulations of stratified drift (or, more appropriately, *washed drift*) fall into two classes which it is important to consider separately because the place of origin and the conditions of deglaciation recorded by each are somewhat different. One class embraces the group of *proglacial deposits,* deposits made beyond the limits of the glacier. Three kinds of deposits constitute this group, accumulated respectively in streams, in lakes, and in the sea.

The other class embraces a group of deposits, some of which have distinctive surface forms, built in immediate contact with wasting ice. These we can refer to collectively as *ice-contact* (*stratified-drift*) features. They include eskers, kame terraces, kames, and features marked by numerous kettles.

To be sure, in many places ice-contact features grade directly into proglacial deposits, so that the distinction between the two classes is somewhat arbitrary. However, the general conditions of origin of the two groups differ so conspicuously that the groups should be thought of as separate.

PROGLACIAL STRATIFIED DRIFT

OUTWASH[1]

Outwash Sediments

In accordance with the distinction just emphasized, outwash is the term used for stratified drift that is stream built—"washed out"—*beyond* the glacier itself. As a sediment most outwash resembles any other deposit made by a stream so heavily loaded that it is continually dropping out more sediment than it picks up. That is to say, outwash is arranged in thin courses of foreset beds, none of which has great continuity because each is partly cut away by younger beds. Variations in grain size are sharp and numerous both horizontally and vertically. Ordinarily there is a very wide range of grain sizes, from good-sized boulders down to fine silt. This reflects the great range of grain size, both in till and in drift in transport in and on the ice from which the outwash is derived.

[1] See E. W. Shaw 1911; M. L. Fuller 1914, pp. 36–38; MacClintock 1922; Woodworth 1897; Flint 1936.

The streams that eventually build outwash do not ordinarily come into existence at the glacier terminus. On the contrary they begin as far back on the glacier as the névé line, the upper limit of melting. From this line on down to the terminus, a distance that may reach many miles, melting of the top and flanks of the ice leads to dozens and even hundreds of little streams. In high latitudes these streams flow only in the summer season and only in the daytime; in lower latitudes they are more persistent. But, large or small, intermittent or otherwise, they find their way downward toward the base of the ice and emerge at the terminus chiefly from the base and along the lateral margins of the glacier. Within the terminal zone of the glacier they have picked up a load of sediment. In a valley with a very steep gradient the meltwater can sluice away all the sediment it had acquired while flowing through the terminal zone. But ordinarily the gradient is insufficient, the coarser part of the load is dropped, and aggradation begins. This interpretation is demonstrated by the fact that the upstream parts of some outwash bodies reach hundreds of feet in thickness.

Aggradation occurs largely because the streams flowing through the marginal zone of the glacier have steep gradients or very efficient channels or both, whereas the gradients beyond the ice are very often smaller, and the proglacial ("beyond-the-ice") stream channels are conspicuously inefficient, being very broad and very shallow.

The average grain size of outwash diminishes downstream, while the degree of roundness rapidly increases. Smoothing and rounding of the larger pieces take place even within the terminal zone, so that by the time the streams have reached the ice terminus some of their coarser deposits are already partly rounded. The facets and striations characteristic of some (though by no means all) of the stones in the basal drift and in the till are quickly worn away,[2] for such features are very rare in outwash.

As outwash is traced downstream past the mouths of tributaries that did not themselves ever carry glacial meltwater, it is diluted by increments of more and more nonglacial alluvium.[3] In a very long stream the outwash is eventually "lost" in this alluvium. But in rare streams like the Mississippi River, which at the maximum of the Wisconsin age drained an 1800-mile sector of the Laurentide Ice Sheet, the massive body of Wisconsin outwash can be traced down to the delta at the river's mouth, a

[2] Wentworth (1923, p. 114) inferred from experimental evidence that under favorable conditions striations could be removed from pebbles in only one-third mile of stream transport.

[3] Very thick and widespread alluvium, mainly of nonglacial origin, mingled with outwash or accumulated against the margin of a glacier have been referred to as *inwash*. This term was adopted in the Glacial Map of North America (Flint and others 1945) to describe the great accumulation of sediments of this origin in eastern Nebraska.

distance of 500 airline miles from the mouth of the Ohio, a principal outwash-carrying tributary, and 700 miles from the position of the nearest glacier terminus. At this distance the outwash sediment consists chiefly of fine sand and silt.

However, erratic pieces of much larger size occur in outwash.[4] They are believed to have been rafted downstream on pieces of ice, either glacier ice or, more often, river ice carried away during the spring thaw and breakup. No very thick piece of ice could travel far down the shallow water of a typical proglacial stream that is actively building outwash. However, as McGee once pointed out, a rafted boulder 4 by 6 by 7 feet, found in the Chesapeake Bay region, could have been floated on a slab of ice 34 by 40 by 2.5 feet — the greatest thickness observed in river ice there today.[5]

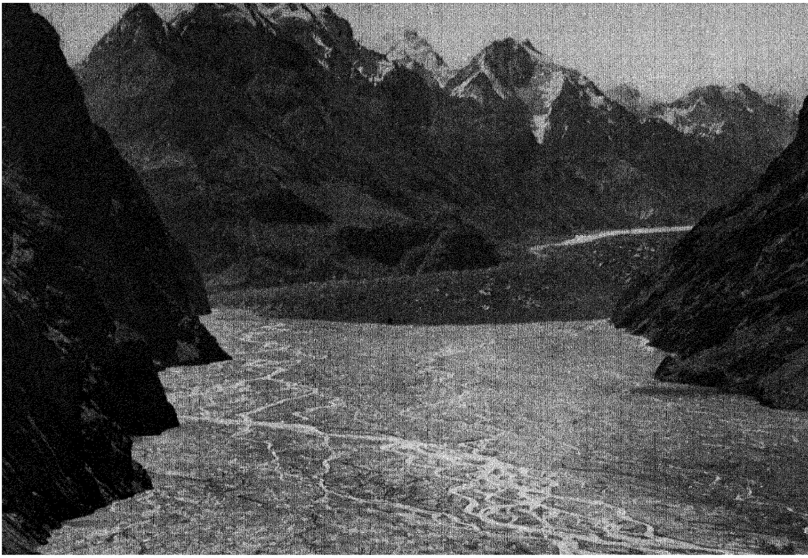
Surface Form

The simplest form taken by a mass of outwash is that of a single fan resting against the terminus of the glacier and with its apex at the point of emergence from the ice of a subglacial, englacial, or superglacial stream. Small fans, some of them hardly more than an acre in extent, are present on the slopes of some large end moraines. Other, larger fans head in pronounced gaps in the moraine. More commonly the outwash mass consists of a row of coalescent fans whose individual apices are still recognizable after outwash deposition has ceased. The massive outwash bodies of Long Island and Cape Cod have this form. Where both meltwater and drift are very abundant the outwash is built up so that it extends gradually headward over the terminus and lateral margins of the glacier, merging with ablation moraine in the terminal zone and with kame terraces along the sides of the ice tongue. There is thus a complete gradation from small, distinct outwash fans to a great mass of outwash that may bury not only any end moraine that may be present but also the terminal zone of the glacier itself through several miles upstream from the actual terminus of the ice.

A long narrow body of outwash confined within a valley is commonly termed a *valley train* (Fig. 37). Both the valley-train and the fan forms of outwash have braided stream patterns while active deposition is in progress. The braided pattern characterizes all streams, glacial or nonglacial, whose bed loads reach the limit of carrying capacity of the streams. Gradients are steep. Gradients of 20 to 50 feet per mile in the

[4] Matthes found erratic cobbles and boulders in the Mississippi Valley outwash apparently more than 200 miles south of the nearest contemporary glacier ice.

[5] McGee 1888, p. 605.



W. Rickmer Rickmers

FIG. 37. Valley train with braided channel pattern. View upstream to the terminal zone of the glacier, blanketed with ablation moraine. The principal streams emerging from the ice are lateral. Fedtchenko Glacier, Pamir Mountains, Tadzhik S.S.R.

headward parts of valley trains are common, and exceptional gradients as steep as 375 feet per mile have been observed.

The filling of a valley with outwash takes place more rapidly near the center than along the sides. Accordingly an active valley train reaches its greatest height along its middle line and slopes slightly toward its lateral margin as well as down the valley. Further, outwash is built up in a main valley faster than most large tributaries, not fed by meltwater, can aggrade their valleys. In consequence the outwash is built into the mouths of tributaries with fan- or delta-like forms. Many tributary streams with small discharge are dammed and ponded, some of them deeply, by the fills at their mouths.[6] The gradual building-up of outwash in some large valleys has filled them to depths of many hundreds of feet. The accompanying rise of the stream profile often leads to diversion of the stream across bedrock spurs or even larger divides. By this method the proglacial Mississippi River was diverted permanently into the valley of the lower Ohio River.

The upstream parts of some outwash masses are pitted with *kettles*—closed depressions created by the melting-out of buried or partly buried blocks of ice after sedimentation had ceased at the site of the kettle. Small kettles in outwash may be attributable to pieces of ice floated away from

[6] E. W. Shaw 1911; 1916; Flint 1936, pp. 1859, 1865, 1876; Thornbury 1944.

the glacier by the proglacial stream. All such streams, however, are shallow, and, as any piece of ice when afloat is largely submerged, no very large depression could be made by ice floated into place. The larger kettles in outwash result from the gradual migration of the head of the outwash upstream over a thin irregular terminal zone of the glacier, which subsequently melts out. The process is discussed later in this chapter. Aside from trenching by major streams, most outwash is long enduring because being very permeable it resists erosion by local runoff.

Most outwash masses are terraced. Terracing occurs in many places as a result of shrinkage of the glacier at the head of the outwash mass. The shrinkage creates a settling basin in which the sediment brought from upstream is deposited. The overflow, passing over the older outwash mass, is thus underloaded, so that instead of adding to the outwash it trenches it. A valley train thus trenched is left as a pair of terraces along the valley sides. A nonglacial stream reoccupying a valley filled with outwash carves the outwash mass into nonpaired stream terraces with decreasing gradient from the highest and earliest to the lowest and latest.

Repeated glaciation of a valley results in the building of successive outwash fills, two or more of which can be recognized in many valleys. The valleys of the Alps are described as carrying remnants of repeated fills of this kind.

Time of Outwash Deposition

The bulk of the outwash visible today was built during the shrinkage of the glaciers of the Fourth Glacial age. Naturally whatever outwash was built during the earlier expansion of these same glaciers was overridden and covered up or destroyed by the ice that flowed over it. However, there is reason to believe that throughout much of the glaciated region less outwash was built during expansion than during shrinkage. The lower regional temperatures, the rapid rate of flow, and the high steep termini of the actively expanding glaciers would have provided a zone of ablation of minimum width, with a correspondingly small discharge of meltwater capable of picking up and reworking the drift contained in the ice. On the other hand the higher temperatures that ultimately caused shrinkage, and the gradual thinning of the glaciers as shrinkage proceeded, and the reduced rate of flow of the ice must have conspired to create a much larger discharge of meltwater year by year. Within the belt of conspicuous drift that forms the marginal part of the region covered by the Laurentide Ice Sheet and within the corresponding belt abandoned by the Scandinavian Ice Sheet, the inner, younger part seems to contain more outwash and other stratified drift than the outer, older part. This fact seems to support the deduction drawn above.

Relation of Outwash to Character of the Till

The belief has been stated frequently that, in those districts in which outwash is scanty in amount, the glacier ice from which it was derived must have wasted largely by evaporation rather than by melting. Aside from the improbability that melting is grossly subordinate to evaporation in any large glacier, even the driest of the glaciated regions (such as southern Saskatchewan and Alberta and western North Dakota and Montana), no matter how poor in outwash, have channels cut by streams of glacial meltwater. In many districts these are large and numerous, and record streams with large discharge. It is probable, indeed, that the now-dry Plains region was much less dry at those times when the edge of the former ice sheet was near by, creating pronounced storminess accompanied by overcasts and precipitation.

More significant still, most if not all of the outwash-poor districts are characterized by till rich in clay and poor in stones. There was simply nothing in these districts from which bulky outwash could be made. The few stones in the drift had to suffice, because the clay and silt, being easily carried in suspension in the meltwater streams, were largely exported. Probably much of this fine sediment now lies on the sea floor.

On the other hand there is some truth in the statement that bulk of outwash is related to climate and therefore presumably to stream volume. On the northeastern (landward) side of the Coastal Mountains of Alaska the end moraines are notably bulkier than those on the seaward side of the same mountains, where temperatures are lower and precipitation is much more abundant. A number of factors are involved, but the chief one may be the greater volume of meltwater on the seaward side of the mountains, which washed away so much drift from the terminal zones of the glaciers that less drift was available for building end moraines.

Compaction

Although, as has been stated, an active valley train is a little higher at its center than at its sides, the surfaces of many thick outwash fills long abandoned by meltwater streams slope gently from sides to center, reversing their presumed original slopes. Where a thick fill is built into and up a tributary valley, its surface may slope back toward the main mass of outwash despite the fact that its stratification clearly shows that formerly it must have sloped in the opposite direction. Such reversals of slope are believed to be the result of differential compaction of the sediments that constitute the fill. When first deposited the sediments have a porosity of 25 to 50 per cent, but this is gradually reduced throughout the entire mass as settling slowly takes place. If the fill is thick and if the

bedrock floor beneath it has a conspicuous slope in some direction (as from the side to the center of a valley), the net compaction will be greatest where the floor is lowest. In this way an initial slope can be reversed through distances up to several miles and may slope backward at a rate as great as 25 feet per mile.[7]

Lake and Marine Deposits

Lake Basins

The distribution and the internal character of the stratified drift shows that a not inconsiderable part of it was deposited not by streams but in temporary lakes held in basins one side of which was formed by glacier ice. The majority of these basins were formed by the terminus or margin of a glacier resting on ice-free ground that sloped down toward the glacier. This situation occurred commonly in mountain valleys[8] and on broad regional slopes. Perhaps the grandest example on record is that of the glacial Great Lakes during the Mankato sub-age. On a front more than a thousand miles long from the upper St. Lawrence to Saskatchewan, great lakes fringed the shrinking Laurentide Ice Sheet. Throughout much of this distance the waters at their northern shores lapped against glacier ice from which bergs broke off and floated southward. Fed largely by meltwater the lakes overflowed through a varied succession of spillways across the divide between valleys draining toward the ice and those draining directly southward and eastward to the sea. The sequence of lakes and spillways is described more fully in Chapter 13.

Another type of basin, smaller and less common, was created by a dam of drift or other deposits across a valley fed with meltwater by a glacier upstream. A conspicuous example is the Hartford Lake which stood in the Connecticut Valley in Massachusetts and Connecticut at some time before the Mankato sub-age, was at least 50 miles long, and endured through hundreds if not thousands of years.[9] The dam was a thick mass of outwash heading at Rocky Hill, Connecticut. The outlet was a spillway over a bedrock threshold several miles to one side of the dam; the relation of these features was like that in many artificial reservoirs: a high earth dike and a lower concrete spillway to one side. The outflow from the lake poured over the ready-made spillway, detouring the outwash-filled part of the Connecticut Valley, until the slow processes of local erosion succeeded in breaching the outwash fill. The outlet then shifted

[7] Flint 1936, pp. 1858, 1861. See also Leverett 1908, p. 5.

[8] See the classic paper on Glen Roy, Scotland, by T. F. Jamieson (1863); also a discussion of glacial Lake Missoula, Montana, by W. M. Davis (1920, p. 135).

[9] Flint 1933.

to this breach and the lake was quickly drained. While the lake was in existence, however, thick deposits of sand, silt, and clay accumulated in it, and they were left as terraces when the Connecticut River, replacing the lake, rapidly eroded them.

Almost identical relations characterize the basins of Okanagan and Skaha lakes occupying the Okanagan Valley in southern British Columbia. The drift dam and the abandoned bedrock spillway near the town of Oliver, and extensive deposits of laminated silt made in the glacial lake itself, at least 75 miles long, are clearly evident.[10]

Still another sort of basin is made by a valley glacier flowing out of a tributary valley into a main valley which it blocks completely. The main valley is then filled with meltwater coming from a second glacier farther upstream. These relations were very common in the North American Cordillera during the Wisconsin Glacial age. An example of which many details are known is the Spokane River Valley in northern Washington with its westward continuation, the Columbia River Valley, as far downstream as Coulee Dam.[11] During the shrinkage of the glacier ice in Wisconsin time the Spokane-Columbia Valley was blocked and its meltwater stream ponded by at least two distinct glacier lobes. One, the Columbia lobe, came down the Columbia past the mouth of the Spokane. The other, the Okanogan lobe, came up the Columbia to the site of Coulee Dam. Both formed lakes, which were partly filled with fine sediment from the glaciers farther upstream. A smaller lobe, midway between the other two, came down the Sanpoil Valley and entered the ponded Columbia, but it apparently did not form a dam. The outlet of the lake formed by the Columbia lobe, which was relatively short-lived, apparently was over the glacier itself, because no abandoned spillway detouring the ice has been found. The outlet of the lower lake was through the spectacular Grand Coulee, a ready-made stream channel that detoured the entire Okanogan lobe.[12] The lake was at least 50 miles long. At its head was built a great delta-like fill of gravel and sand.

A large number of glacial lakes, usually small and inconspicuous, are created where a valley glacier in a main valley forms a dam across the mouth of a tributary valley and ponds the water in it. Nearly every glaciated region of considerable relief shows examples of this kind of lake.

Glacial-Lake Deposits

The deposits made in glacial lakes are of four principal kinds: deltas, bottom deposits, rafted erratics, and shore features.

[10] Flint 1935*c*.
[11] Flint 1936, pl. 6.
[12] Flint and Irwin 1939.

Naturally a delta can be built into a lake by any stream entering from the land. Our interest is primarily in deltas built by streams entering from glacier ice. They correspond to outwash fans but differ from them in that they are built into standing water. Deltas that are essentially parts of end moraines, in the Salpausselkä ridges of Finland, were described in Chapter 7. Some of these, as well as many other deltas built from the ice, have their origin in eskers as described beyond in the present chapter. These features are known as *esker deltas.*[13] Ordinarily their upstream parts are built against or upon the glacier terminus. When the ice melts the sediments slump away, altering the form of the delta. Aside from this and from the presence here and there of depressions left by the melting of ice blocks, esker deltas differ very little from ordinary deltas. Their sediments grade rapidly outward from coarse to fine, and their foreset slopes merge imperceptibly into the faintly undulating lake floor.

The sediments of the lake floor consist of the finer products of glacial erosion, chiefly silt and clay, whereas most esker deltas consist mainly of gravel and sand. A small delta can be made very quickly, perhaps in a single season, but in this time only a very thin deposit of fine sediment can be spread over the floor of an extensive lake. Many glacial lakes are drained, and deposition in them is thereby brought to an end, before more than a thin deposit of sediment has accumulated on their floors. Others, particularly small lakes, have been completely filled with sediment. In these the deposits become coarser from base to top, indicating gradual shoaling of the water and correspondingly increased capacity of the lake currents to sweep sediment in suspension toward the point of outlet.

The fine-grained bottom sediments of many glacial lakes occur in pairs of coarse and fine layers called *varves,* which are believed to represent annual cycles of deposition and are therefore used as a basis for determining the duration of the lake and for other time calculations. The significance of varves is discussed in Chapter 18.

The bottom sediments of many glacial lakes include erratic stones and boulders both singly and in clusters. These apparently were rafted down the lake on floating ice, either bergs broken off from the glacier or lake ice broken up and set free during the spring thaw. In some sections of lake-floor deposits these erratics diminish from the base upward, suggesting that floating ice was common at first but later disappeared as the glacier shrank away and as the climate became warmer. The deposits of the earlier glacial Great Lakes include "nests" of erratic boulders and stones, each "nest" apparently marking the site of an overturned or

[13] Tarr 1909*b*, p. 23. They have also been called sand plains (W. M. Davis 1890). For a good description see Trotter 1929, p. 573.

grounded berg which left its load on the spot.[14] Till-like deposits lying upon or between lake sediments probably originated in overloaded bergs which sank to the lake floor, carrying their heavy loads with them.[15]

It is worth noting that sizable bergs can float away only where the glacier terminates in comparatively deep water. A glacier aground on its own delta, or otherwise ending in very shallow water, can not discharge any but very small pieces of ice.

Large glacial lakes such as the glacial Great Lakes are marked with wave-cut cliffs, beach ridges, spits, bars, and similar features along their shores. As such features are directly or indirectly the result of wave work they are best developed in wide lakes where large waves could be generated and where there was ample time for them to work. Around the shores of many former glacial lakes that were small or at least narrow no such shore features are present. This is a result of several factors among which are (1) feeble waves and currents, (2) short life of lake, (3) fluctuation of water level by changes in the outlet or by crustal warping, (4) susceptibility of faint shore features to destruction by mass-wasting, (5) presence of bare firm bedrock or glacier ice along parts of shores.

Glacial-Marine Deposits

The principal difference between glacial-marine deposits and glacial-lake deposits, besides the obvious differences in fauna and flora and therefore in fossil content, is that varves do not form in salt water. Sea-water is an electrolyte which causes the clay in suspension to flocculate and settle along with the silt. In fresh water, on the contrary, the silt settles out quickly during the great summer discharge of meltwater, leaving the clay to settle slowly in autumn and winter after the lake has frozen over.[16] No varved sediments therefore are marine.

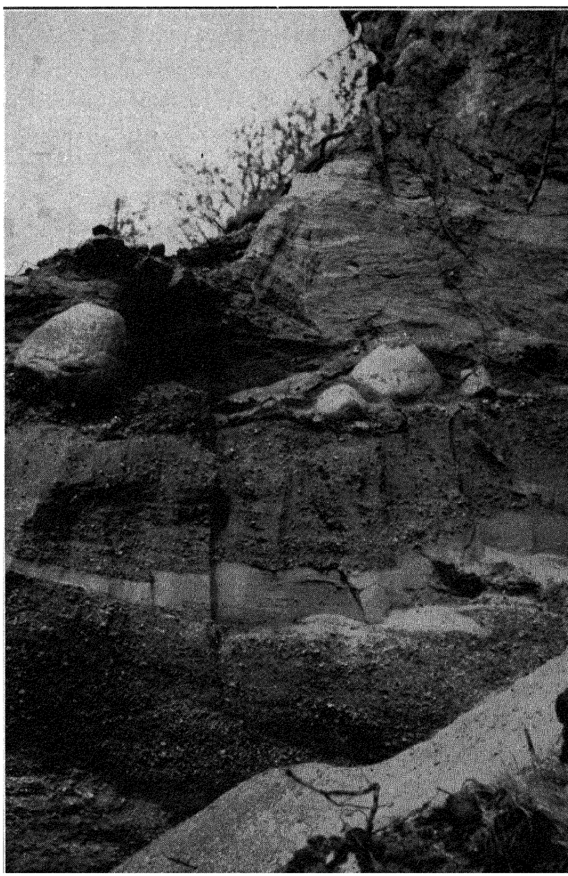
The details of deltas and similar features built where glacier termini stood in the sea are not likely to have been preserved except in protected bays because they were open to destruction by strong wave action. The irregular form of many such marine deposits suggests that they were made in openings between masses of stagnant ice which protected the sediments from wave erosion until crustal upwarping had lifted them above sealevel out of reach of the waves.[17]

[14] See von Engeln 1918; Kindle 1924.
[15] Cf. Cooper 1935, p. 30.
[16] Antevs 1925*b*, pp. 33–44; Douglas Johnson 1939, pp. 43–48.
[17] Cf. Trefethen and Harris 1940; see also Katz and Keith 1917.

ICE-CONTACT STRATIFIED DRIFT

Characteristics of Ice-Contact Sediments

The sediments of ice-contact stratified drift have three general characteristics that distinguish them from proglacial sediments: extreme range



Henry Retzek

Fig. 38. Detail exposure of ice-contact sediments, Hawk Creek, Minnesota, showing three typical features: (1) extreme range and abrupt changes of grain size; (2) interbedded till (at top of section); (3) deformation (at base of section).

and frequent and abrupt changes in grain size; intimate association with till; and deformation. It might be added that fewer individual pieces are stream worn than in proglacial deposits.

Figure 38 illustrates all these characteristics, each of which in its own

way reflects the immediate proximity of glacier ice during deposition. Whether accumulation takes place upon, against, or underneath the wasting terminal zone of the glacier, it is likely to be sporadic and irregular, with no intervening distance to smooth out the differences between summer and winter, noon and midnight, in rate of melting and the release of sediments. The same site may successively see a rushing stream, a quiet pool, a small avalanche of boulders, and actual overriding by slowly flowing ice, folding or faulting the layers of sediment or smearing till upon them. Ice is likely to melt out from beneath the accumulating sediment, or from a supporting position beside it, causing sagging, collapse, slumping, land-slipping, earthflow, and a variety of other distortions and disturbances. In such a place, in short, anything can happen, and it usually does.[18]

Surface Expression of Ice-Contact Features

The masses of stratified drift that have the peculiar characteristics just described also have a surface expression that is distinctive when not subsequently altered by mass-wasting and other kinds of erosion. Surfaces are either constructional or the result of deformation through the melting of supporting ice.

The typical constructional surface is the sideslope or face of a mass of sediment that was built up against a steep supporting wall of glacier ice. The ice has melted away and the adjacent part of the sedimentary mass has slumped down. Although slumping destroys the actual surface of contact with the former ice in detail, the larger features of the surface of contact are very commonly preserved. Thus indentations remain where great protuberances in the ice once stood, and projections, mostly cuspate, remain where the ice was marked by crevasselike re-entrants. Such features as these characterize the faces of kame terraces, described beyond, and isolated masses of sediment accumulated in openings surrounded by thin wasting ice. They were long ago aptly termed *ice contacts,*[19] and if this term is used broadly and not too strictly it is applicable and therefore useful.

A second type of constructional surface consists of the outer part of a body of stratified drift that was built up beneath an overhanging wall of ice. Depressions in such a surface may be the direct casts of downward projections of the ice. Although individual surfaces of this kind are not

[18] These sedimentary and structural characteristics taken together have been collectively referred to as "kame structure." As this is a poor term from every point of view it is not used here.

[19] Woodworth 1899. Recently formed ice contacts are well described in Tarr 1909*a*, pp. 87, 96, 98, 110.

likely to be extensive, the type is probably not uncommon in places where glacier margins were bathed in standing water. This is because convection currents in the water tend to undermine the ice by melting and thus to create an overhanging wall.

Caution is necessary in interpreting a steep hillslope marked by knolls and closed depressions as an ice contact. Landslips on steep slopes, particularly ordinary river bluffs and stream terraces, produce such features and thus superficially resemble ice contacts. Detailed examination is required in order to establish their true origin.

R. F. Flint

FIG. 39. Section of ice-contact laminated silt on Columbia River near Bossburg, Washington. Successive bodies of silt each many feet thick and each deposited in a quiet pool have collapsed, mostly with accompanying contortion, as supporting ice melted away.

In addition to the constructional surfaces there is the surface that results from gradual collapse of an originally more or less flat surface of stratified drift as ice melts out from beneath it. This happens at the heads of massive bodies of outwash that were gradually extended headward over the thin terminal part of the glacier. As the ice melted, following the close of outwash deposition, the superglacial part of the outwash gradually sank and settled—collapsed—onto the ground beneath (Fig. 39). The amount of settling was in direct proportion to the thickness of the ice, which increased headward as the thickness of the outwash decreased. The general surface slope was thereby reversed, but in addi-

tion sags, swells, and unsystematic irregularities were developed in the surface detail, partly by reason of variations in the thickness of the formerly underlying ice. Broadly speaking, collapsed surfaces develop where stratified drift was comparatively thick and where the underlying ice was comparatively thin. In contrast ablation moraine, which is fundamentally a collapsed deposit, is a product of drift (mostly unstratified) that was thin in relation to the ice upon which it formed.

Principal Types of Ice-Contact Features

Kame Terraces

In 1874 Jamieson[20] interpreted correctly the origin of terraces with ice-contact faces along the sides of valleys in the Scottish Highlands. It

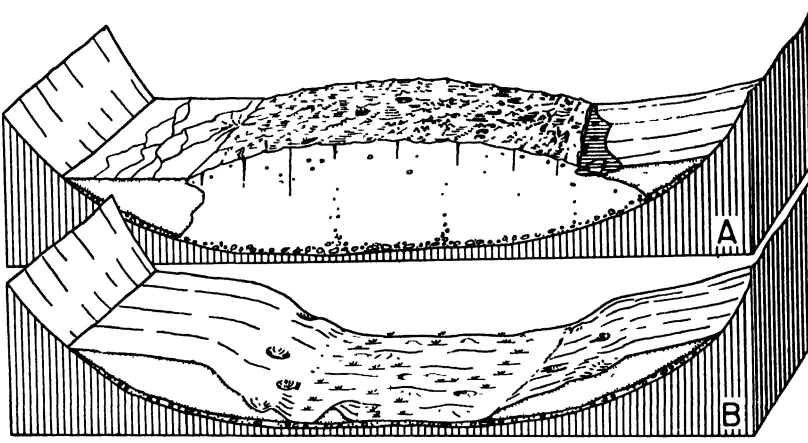


Fig. 40. Ideal kame terraces under construction (*A*) and after deglaciation (*B*).

was not until nineteen years later that a description of the actual formation of stream deposits along the sides of an existing glacier appeared in print.[21] In that same year Salisbury[22] used for such features the name *kame terrace*. This feature is now widely recognized as an accumulation of stratified drift laid down chiefly by streams between a glacier and an adjacent valley wall and left as a constructional terrace after disappearance of the glacier (Fig. 40). The significant distinction between kame terraces and ordinary stream terraces is that they are not the remnants of a mass of sediment that formerly filled the valley from side to side but were never much more extensive than now.

[20] Jamieson 1874.
[21] I. C. Russell 1893, p. 236.
[22] Salisbury 1893*b*, p. 157.

Usually narrow, a kame terrace has a comparatively flat top and a down-valley profile like that of a valley train. If very narrow it may be discontinuous where bedrock spurs and bosses interfered with deposition or where the depositing stream passed over ice for a short stretch. The sediments composing it were derived chiefly from the drift in the former adjacent glacier and secondarily from tributary streams and from the valley side. The terrace face initially has an ice-contact form, though in many places this is later destroyed as the terrace is undercut by the post-glacial stream that succeeds the glacier. The terrace top is likely to be pitted in places by kettles. These are more numerous near the face of the terrace, the most recently built part, than near its inner edge where buried ice is more likely to have melted out before stream deposition ceased, so that the early-formed kettles were filled with sediment.

Many of the more massive kame terraces grade down the valley into valley trains. Through the terraces the outwash is virtually continued headward along the sides of the valley. But in the central part of the valley the outwash has collapsed to form an irregular reversed slope. At some places[23] several kame terraces occur in series, one below the other, marking the successive positions of ice-marginal streams as the former glacier thinned during its wasting. Many a terrace of this kind can be traced downstream along the valley side into a saddle or notch in a bed-rock spur, where it ends at the level of the notch. The bedrock floor in these places acted as baselevels for the streams that built the terraces.

Most kame terraces are very short. Even the longest hardly exceed a maximum length of a few miles. In theory a terrace could extend head-ward along a glacier to the névé line, but no farther, because above this line there is no meltwater. In actual fact probably no terrace extended even this far.

Kames[24]

Another common group of ice-contact features consists of kames[25] which we may describe as low, steep-sided hills with a knoll-like, short-ridge-like, or mesa-like form, and consisting of ice-contact stratified drift. The ridge-like variety have been called crevasse fillings.[26] Kames originate in at least two principal ways. Some are bodies of sediment deposited in crevasses and other openings in or on the surface of stagnant or nearly stagnant ice which later melted away leaving the accumulated sediment

[23] A good description is given in Daly 1912, p. 590.

[24] See Woodworth 1899; Fairchild 1896; Cook 1943.

[25] *Kame*, a term of uncertain origin first used in Great Britain, has been employed in many different senses. While it awaits formal redefinition it is used here in a restricted sense believed to coincide with the views of most American glacial geologists.

[26] Cf. Flint 1928.

in the form of isolated or semi-isolated mounds. Kames of this kind are common at and just beyond the faces of kame terraces. Had the terraces grown wider these kames would have been incorporated in them. Another type of kame consists of small deltas or outwash cones built out from the ice which later melted, collapsing and isolating the mass of sediment to form an irregular mound.[27]

These examples and their implications make it clear that kames grade into kame terraces, collapsed masses, some types of ablation moraine, and some eskers, and also form integral parts of some end moraines. The majority of kames are intimately associated with kettles and, with them, record thin, crevassed, and stagnant ice.[28] Some of them form a reticulated pattern on a large scale, suggesting a fracture pattern in the former ice.[29]

Kettles[30]

A *kettle* is a depression that occurs in drift, usually stratified, and that has been made by the wasting away of a mass of ice that had lain wholly or partly buried in the drift. Although depressions of other kinds occur in drift, they are not kettles.

The majority of kettles are small. Few exceed a mile in greatest diameter, though some of the largest kettles in north-central Minnesota attain long diameters of at least 8 miles. Few kettles are more than 20 or 30 feet deep, though depths as great as 150 feet have been reported. Though they may have any shape, they tend to approach the circular — the shape any separate ice mass will tend to assume as wasting progresses. Kettles occur singly, in groups (especially linear groups), or in such profusion that the body of drift in which they occur appears a maze of hillocks and depressions sometimes described as "kame-and-kettle topography."

Kettles occur in stratified drift and much less commonly in till. Those in till are likely to be irregular in form; it is particularly difficult to differentiate these from depressions having other origins.

At least three common types can be recognized (Fig. 41). The largest and most conspicuous kettles result from the melting of relatively thick projecting ice masses. They have steep sideslopes caused by sliding of the sediment when the ice support is removed. Thinner, buried ice masses produce shallow kettles formed by collapse rather than by sliding.

In a broad sense kettles are the counterparts of kames. In general both features are the product of the wasting of thin glacier ice detached from the main body of the glacier and therefore stagnant. The fact that the

[27] M. L. Fuller 1914, p. 32.
[28] Flint and Demorest 1942, pp. 126–128.
[29] Richter 1937, pp. 157, 158, 164.
[30] M. L. Fuller 1914, pp. 38–44.

deepest kettles are no deeper than the depth of crevasses (about 150 feet) indicates that kettle-forming ice masses were so thin that the zone of flow had already been pinched out and the base of the zone of fracture had reached the base of the ice, as set forth in Chapter 3.[31]

Kettles are peculiar to the terminal zone of a glacier where thinning is actively in progress. Usually meltwater streams build sediment around and between wasting masses of ice, the remnants of the terminal zone of an earlier phase of deglaciation. Beyond these masses the sediments take the form of outwash.

How long an ice block can last before wasting completely away is not known. Under most conditions ice exposed at the surface wastes rapidly, but ice buried beneath even a comparatively thin mantle of sediment, especially if the sediment is not permeable so that water circulation is

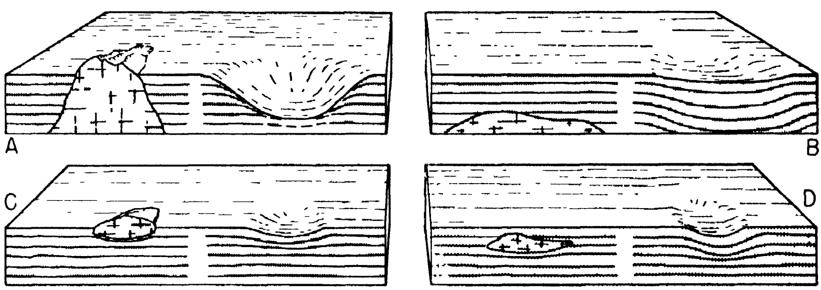


FIG. 41. Three common types of kettles and their inferred origin, idealized. (In part after M. L. Fuller 1914.)

A. Projecting ice mass and kettle formed from it.
B. Buried ice mass and kettle formed from it.
C, D. Floated-in or dropped-in ice masses and kettles formed from them.

inhibited, probably wastes slowly because loose rock is a very poor conductor of heat. The annual sedimentary layers laid down in a Swedish lake overlying buried ice, and later collapsed, number 70; hence this buried ice required at least 70 years to waste out.[32] It has been thought that ice found buried beneath sedimentary layers in Arctic regions now free of glaciers was "fossil" glacier ice inherited from a colder time in the latest glacial age. However, most if not all such occurrences are now recognized as ground ice formed in place probably in comparatively recent time.

[31] Flint and Demorest 1942, p. 125.
[32] De Geer 1940, p. 47.

Eskers[33]

The most remarkable form of ice-contact stratified drift is the *esker,*[34] a long, narrow ice-contact ridge commonly sinuous, and composed chiefly of stratified drift.

Eskers range in height from a few feet to 50 or even more than 100 feet, in breadth from a few tens to a few hundreds of feet, and in length from a fraction of a mile up to nearly 150 miles if gaps, which occur in every long esker, are included. Sides are generally steep, approaching the angle of rest of the esker sediments. Crests are smooth or broadly hummocky. Kettles pit the broader parts of some esker tops.

No esker is straight. All are sinuous, through nearly the same range as normal streams. That is, some are only slightly sinuous whereas others have great curves resembling meanders. Many eskers, including most long ones, have tributaries, forming with them patterns like those of streams. A few appear to have poorly developed distributaries.

Most eskers occur in regions of rather low relief, and in broad outline their trend parallels the direction of flow of the latest of the former glaciers. Where conspicuous valleys are present, some eskers follow the floors of the valleys, but not where valley trends diverge greatly from the direction of former glacier flow. Esker gradients are gentle, but they are not necessarily continuous. That is to say, an esker may climb *up* a valley for several miles, pass over a low divide, and descend the far slope. Crossings of divides are invariably at low saddles or gaps, and at such a crossing the esker is likely to be notably discontinuous.

Some eskers merge downstream into deltas, groups of kames, or outwash masses, but many end without apparent relation to other features.

Sand and gravel are the chief constitutents of most eskers, though both silt and boulders are present in places. Consequently eskers are a source of material for road surfacing, and the resulting pits expose their internal character. In most places the sediments are crossbedded as though deposited by streams. The general direction of the foreset layers and the consistent decrease in grain size indicate the direction in which the stream flowed. The crossbedding, however, is much less regular and even than in outwash; lenses and pockets of till and only slightly washed drift are present. The sediments coarser than sand are generally waterworn.

Regardless of the stratification in the central part of an esker section,

[33] Good general references include W. O. Crosby 1902; Woodworth 1894; Chadwick 1928; Giles 1918.

[34] This term, derived from a Gaelic word, was first used in a technical sense by Close (1867) in Ireland. The word *ås* (pl. *åsar*) is widely used in Scandinavia for these features, and in England the anglicized word *ose* is used by some.

the material at the sides tends to lie parallel with the steep sideslopes. This arrangement results in part from sliding of surface sediments after they had been deposited in more nearly horizontal positions. In some eskers at least it is also in part the result of original deposition at angles oblique to the stratification of the central bulk of the esker. Cross-sectional exposures of these eskers have a false appearance of possessing anticlinal structure.

The sediments of which an esker is built are closely similar to the constituents of the till in the vicinity. The similarity suggests that they have a common origin in the drift carried in the ice. A very detailed study of an esker north of Åbo in southwestern Finland led to some interesting results.[35] For 18 miles this esker lies on a single type of bedrock covered with till of somewhat variable composition. The pebbles in the esker had clearly been derived from the till and on the average had traveled less than 3 miles from their former positions in the till. Comparable results were obtained from study of an esker in the Kennebec valley in Maine.[36] These data suggest that eskers are built of local rather than far-traveled sediments.

As Hummel was apparently the first to point out, it is almost universally agreed that eskers are the deposits of glacial streams confined by walls of ice and left as ridges when the ice disappeared.[37] It is further agreed that as the eskers in any district parallel the striations and other records of the *latest* direction of glacier flow there (Fig. 42), and as they are closely related to deltas, kames, and outwash, they are late deposits, having been built in the terminal zone of the glacier shortly before it disappeared from the district. However, not all eskers were built under just the same conditions, although all are constructional rather than erosional features. It is probable, in fact, that eskers are formed in several rather distinct ways, of which two seem more common by far than the others.

The most common mode of origin appears to have been in tunnels (less commonly open canals) at the base of the glacier, during so late a phase in the deglaciation process that the ice was thin and stagnant or nearly so. It is unlikely that tunnels could easily form or, once formed, stay open unless the ice that inclosed them was nearly motionless. If stagnant, the ice must also have been thin. Water derived chiefly from surface melting worked its way downward through crevasses and other openings and at the base of the ice enlarged systems of openings to form tunnels. The lowest possible channelways were sought (which is why

[35] Hellaakoski 1931.
[36] Trefethen and Trefethen 1945.
[37] Hummel 1874.

FIG. 42. Eskers of Fennoscandia. (Compiled from various sources.)

eskers generally occupy valleys). Flowing through these openings, chiefly under hydrostatic pressure, the flowing water emerged in ponded water (a glacial lake or the sea) at the glacier's terminus. There is little in the eskers to indicate whether a long individual was formed at the same time throughout its length, or whether its downstream part was built first and was gradually added to in the upstream direction as thinning and stagnation affected an increasingly wide terminal zone. Unless the whole of the esker, after it was completed, was protected by inclosing ice, it is not easy to account for its preservation from destruction by proglacial stream erosion or from burial beneath outwash.

Perkins[38] observed that the long eskers in Maine have increasing continuity, with fewer gaps, as they are traced upstream, and that in the same direction the proportion of erratic stones in them increases. He suggested that this might mean a stagnant terminal zone narrow at first but increasingly wide as thinning progressed. An essentially similar view was expressed by Andersen.[39]

Another, much less common, way in which eskers originate was first suggested by Shaler[40] and later detailed almost simultaneously by De Geer[41] in Sweden and by Hershey[42] in North America. De Geer showed that certain eskers in Sweden consist of short segments, each segment beginning upstream with coarse gravel and grading downstream into fine sediments. The coarse upstream part is narrow, but downstream the esker broadens into a distinct delta. From these facts he inferred that each segment represented the deposit made during one year (chiefly in the summer melting season). The narrow part of the esker was made in a short subglacial tunnel leading to the glacier terminus, at which the delta was built and beyond which the tunnel stream was free to spread beyond the confining walls of the tunnel. Small eskers in northern England[43] and in southwestern Quebec[44] have a similar character, broadening into deltalike mounds at intervals of a few hundred feet. In the examples cited there is clear evidence that the ice terminus constituted one shore of an extensive glacial lake. All the important conditions favor this mode of origin: the well-defined clifflike terminus characteristic of a glacier margin in standing water, the rapid wastage that occurs under these conditions, and the preserving effect of submergence of the esker segments as fast as they are aban-

38 Leavitt and Perkins 1935, p. 71.
39 Andersen 1931*b*.
40 Shaler 1884.
41 De Geer 1897.
42 Hershey 1897.
43 Trotter 1929, p. 573.
44 Norman 1938.

doned. Probably all clearly segmented eskers were formed in this way, though they constitute a small minority of all the eskers on record. On the assumption that each segment was made in a single year this type of esker affords an interesting possibility for determining the average rate of deglaciation during the period measured by the number of segments in the esker.

Two other hypotheses hold that eskers are deposits in superglacial stream valleys[45] and in englacial tunnels[46] respectively. Both hypotheses state that as the glacier thins the deposited sediments are gradually let down onto the ground beneath. The chief merit claimed for these hypotheses is that they explain the fact that some eskers climb over divides, without the necessity of supposing that the streams were controlled hydrostatically. Against them are two principal facts: (1) Most eskers do not trend indiscriminately across country, as they should do if superposed from upon or within the ice. They are highly selective, following valleys through long distances and crossing divides at conspicuous low points. This could happen only if they were built on the ground, under the guidance of the local topography. Indeed the englacial hypothesis is an attempt at a compromise by keeping the ice tunnel close enough to the ground to be influenced by local topography. (2) The process of superposition involves the process of collapse on a large scale. It is highly improbable that either the surface form or the internal stratification of an esker could be preserved through this process. On the other hand collapse is minimized in the concept that superglacial streams nearly free of sediment cut down through nearly clean ice into the basal drift-rich zone, where they picked up a load and redeposited some of it to form eskers. Some eskers may have originated in this general way.

A curious occurrence near Strö, in Denmark, is that of a smooth-crested esker with no discontinuities, cut through by a deep gap. The gap is obviously stream-made, but it was not made by a modern stream, for the floor of the gap lies above the surrounding country. The esker, built in a subglacial tunnel, was gradually exposed by thinning of the inclosing ice, but, as exposure began, a superglacial stream was superposed across the esker, gradually cutting the gap.[47] The stream was diverted, or ceased to flow, before the ice at the sides of the esker had entirely disappeared.

In an attempt to evaluate the relative merits of these hypotheses Tarr[48] observed that, despite the many superglacial streams and sources

[45] Holst 1876–1877; W. O. Crosby 1902; Tanner 1937; Sproule 1939.
[46] Alden 1924, p. 54; Okko 1945.
[47] S. A. Andersen, *unpublished.*
[48] Tarr 1909*a*, p. 96.

of abundant sediment in the widespread ablation moraines on the stagnant parts of Alaskan coastal glaciers, no superglacial stream deposits had been observed on them. The water soon disappears through some opening to the base of the ice, carrying its load with it. On the other hand he found large streams discharging from the mouths of huge tunnels at the base of the ice, "with great velocity, and heavily charged with sediment, including boulders often up to a hundred pounds in weight. There can be no reasonable doubt that there are eskers forming along the beds of these streams, such as the Kwik River, which has a subglacial course of six miles." Tarr concluded that, under Alaskan conditions at present, superglacial eskers are impossible.

In summary it appears likely that many eskers are subglacial tunnel deposits, and that a few were built headward in successive segments each marked by a delta where the esker stream entered a glacial lake. The former type requires extensive thin stagnant ice;[49] the latter type demands only a narrow zone of such ice, no wider than the length of a single segment of the esker. That some eskers have been let down from an englacial position has been urged by Okko.[50] The case for a superglacial origin has not been well established.

If we take a still broader view and consider the whole assemblage of ice-contact features, notably kames, kame terraces, large and abundant kettles, and small eskers, we find that they are distributed through a much wider belt than the large eskers, but that they are rarely abundant within 100 miles of the drift borders of the Fourth Glacial age. These features indicate thinning, though not through a wide belt. They may be thought of as the forerunners of the great eskers in recording progressive thinning of the outer parts of the great ice sheets.

SUMMARY

The surface form and internal character of stratified drift show that it is of two distinct kinds: proglacial stream, lake, and marine deposits; and ice-contact deposits. The close association of the last with thin stagnant or nearly stagnant ice suggests that ice-contact features are formed most commonly during the later phase of the shrinkage of an ice sheet when the ice has become thin through dwindling nourishment.

[49] Flint and Demorest 1942, p. 130.
[50] Okko 1945.

Chapter 9

EXPANSION AND SHRINKAGE OF GLACIERS INFERRED FROM DRIFT

GENERAL SIGNIFICANCE OF TILL AND OF STRATIFIED DRIFT

Some conclusions as to the general interpretation placed on till and stratified drift are possible, but they must be carefully qualified. The deposition of till is favored both by rapid glacier flow and by low temperature which reduces melting. This combination of conditions is met with during the expansion of a glacier, regardless of its type or location. A maximum proportion of the drift it carries is deposited (chiefly by lodgment) as till, and only a minor proportion is moved by meltwater. Also, part of the stratified drift deposited in front of the expanding glacier is reworked and redeposited as till, further reducing the proportion of stratified material later exposed to view. This last action is reflected in tills that contain a large number of rounded stones.

The deposition of stratified drift, on the other hand, is favored by high temperature (which causes melting) and by slow glacier flow. These conditions prevail during the shrinkage of a glacier. Reduced flow is a direct result of thinning, which in turn is a response to increased temperature. Meltwater not only is abundant but is spread through a terminal zone of greater width, owing to thinning. In consequence much drift that in a more active glacier would be lodged on the subglacial floor as till is seized by meltwater and deposited as both ice-contact and proglacial stratified drift. During a final deglaciation it is not reworked or covered up by subsequent glacial expansion and hence remains in evidence.

The foregoing comparison is subject to this modification, that in regions of fine-grained sedimentary rocks, such as the Great Plains and Central Lowland in North America and the steppes of southern European Russia, the bulk of the glacial sediments are so fine grained that they were largely carried from those regions without being dropped en route. Hence it is not justifiable to conclude that because outwash is scarce there was not much meltwater. In the absence of outwash, the work of meltwater is still clearly demonstrated by stream trenches now dry or showing evidence of former great discharge.

SIGNIFICANCE OF END MORAINE

End moraine (not including the ice-contact head of a massive accumulation of outwash) records actively flowing ice together with at least temporary equilibrium among the factors of nourishment, flow, and ablation. Most end moraine is a ridgelike thickening of ground moraine built by lodgment of drift, and ground moraine of this kind also indicates ice that was actively flowing.

On the other hand the absence of end moraine at the margin of a drift sheet does not necessarily indicate that the glacier ice was not actively flowing at the time the drift was laid down. It may be that the drift in transport was too scanty to build a moraine, that ablation was more than equal to nourishment and flow so that retreat of the glacier margin began as soon as the ice had spread to its outer limit, or that meltwater had sufficient discharge or gradient to carry away all the drift released from the ice.

Drift borders (the outer limits of drift sheets) that are unmarked by end moraine have long been recognized and described as "attenuated" drift borders. In the older drift sheets and in mountain valleys this condition may be in part the result of erosional destruction of formerly existing end moraines. But in large part it appears to result from lack of equilibrium in the glacier. This conclusion is not unnatural, as equilibrium is considerably less likely to occur than not to occur. The chances favor both glacier expansion and glacier shrinkage over glacier equilibrium. Mile for mile, "attenuated" drift borders appear to be at least as extensive as drift borders marked by end moraine, in both North America and Europe.

GROUND- AND END-MORAINE SEQUENCE COMPARED WITH ICE-CONTACT SEQUENCE

It has been stated above that a sequence of glacial-drift deposits consisting chiefly of ground moraine and end moraine built of till by the process of lodgment records the work of actively flowing and therefore comparatively thick ice. In contrast, an ice-contact sequence consisting of collapsed masses, kame terraces, kames, and eskers is produced by ice that is thin and either feebly flowing or stagnant. In some regions ice-contact features occur in broad belts. Although within such a belt it is difficult to discriminate between deposits made successively and those made contemporaneously the existence of such belts suggests (though it does not prove) thinning throughout a broad peripheral zone of the glacier with melting occurring over a wide area as indicated by very bulky coarse sediments. If this condition obtained, it must have resulted

mainly from decrease in nourishment rather than increase in ablation with nourishment unchanged, for, if increase in ablation alone had occurred, it would have been confined to a narrow peripheral belt and widespread thinning would not have occurred.

It may be mentioned parenthetically that marked topographic relief in the surface beneath the ice should hasten the process of ablation and extend the zone of thinning both by furnishing topographic obstacles to ready flow and by reflecting heat on to the ice from nunataks and exposed ridges. It is difficult to assess the relative influence of these factors, but both may have been great.

The types of drift built by the latest, and now vanished, great ice sheets in North America have been so inadequately mapped that firm inferences as to their broad significance are not yet justified. However, there appears to be, on both continents, a zoned arrangement of the drift that is especially conspicuous in central North America and in the Baltic region of Europe. Through an outer belt 200 to 300 miles wide the drift consists chiefly of clay-rich till in the form of ground moraine with many concentrically arranged end moraines. Inside this belt lies a wide zone in which, though till and end moraines are still present, ice-contact features (including small eskers) are prominent. As this belt is followed inward toward the central part of the glaciated region, end moraines become very rare indeed, and great groups of eskers appear which probably were formed within broad areas of contemporaneously thin ice.

From this we may draw the very speculative inference that during the early part of the last deglaciation of North America and Europe (Iowan and Tazewell sub-ages in North America, and Warthe and Brandenburg sub-ages in Europe) the two ice sheets were large and vigorous enough to maintain near-equilibrium, albeit with a net deficit most of the time. Hence significant thinning did not take place and the drift was actively plastered onto the ground and built up to form till and end moraines.

Gradually, however, diminishing supply resulted in increased shrinkage, especially by thinning. The net deficit became acute after the beginning of Cary time in North America and after the building of the Brandenburg moraines in Europe. Ice-contact deposits replaced till increasingly as thinning slowed down rates of flow. Re-expansions of the ice, however, still occurred, as indicated by the Port Huron moraine and other end moraines. Thereafter shrinkage reduced the ice to such small area and thickness that nourishment was drastically curtailed, notable thinning occurred, and eskers could form throughout a belt 100 to 150 miles wide. In New England massive eskers occur nearer to

the outer limit of the glaciated region than elsewhere, perhaps because of greater relief.

If this speculation should prove justified by further and more detailed examination of the glacial drift, the kame terraces, kames, and collapsed masses of the subperipheral belt may be regarded as the forerunners of the great esker systems farther toward the interior. It is only fair to say that the dearth of till and end moraines in the great central areas (Finland and the Canadian Shield) must be in part the result of lack of materials with which to build such features, the rock types in both regions being unfavorable for the production of much drift. However, this condition applies equally well to ice-contact features, so that the general speculation still appears justified.

INFERRED LONG PROFILES OF FORMER GLACIERS

Before concluding this discussion we may draw attention to drift and other features that have been held to indicate the long profiles of the surfaces of terminal parts of former glaciers and thus to reveal glacier thicknesses. Attempts have been made to reconstruct upper-surface profiles from end moraines, upper limits of drift, and ice-marginal channels eroded by meltwater on the slopes of hills and mountains. Even were the profiles thus reconstructed actual rather than minimal (because the glacier may have reached higher than the highest glacial features now visible), and even were the evidence distinct enough for detailed reconstruction, the resulting profiles would not have great value. The profiles of all glaciers are continually changing, and the exact profile at any one time gives little useful information. However, a profile steep enough to imply thick ice gives some suggestion that at that time and place the glacier was vigorously active. Thus figures obtained by Thwaites[1] in the Baraboo district, Wisconsin, on the southwest margin of the Green Bay Lobe of the Laurentide Ice Sheet during the Cary sub-age, indicate slopes as great as 370 feet through the first mile back from the glacier terminus. That the ice was actively moving there and then is shown by an end moraine that marks the margin of the vanished glacier.

[1] F. T. Thwaites, *unpublished.*

Chapter 10

DRAINAGE, ICE-THRUST FEATURES, AND EOLIAN DEPOSITS

GLACIAL DRAINAGE[1]

Subglacial Drainage

Meltwater may form anywhere on the glacier surface below the névé line. Flowing away as intermittent streams, it forms drainage systems that include superglacial, englacial, subglacial, marginal, and proglacial segments. After the glacier has disappeared, the first two leave few traces or none at all, but the other three leave traces from which some inferences as to the former drainage can be drawn.

Small streams cascading down through crevasses and other openings onto firm bedrock in the subglacial floor cut potholes.[2] The positions of the potholes on the tops and sides of hills preclude their having been formed by normal streams unhampered by walls of ice. Still smaller streams falling similarly onto the tops and steep slopes of subglacial rock hills wash them clean of their till cover and channel the rock.[3]

Sizable streams flowing beneath the ice and not loaded with coarse sediment cut shallow trenches in the till and other loose material at the surface. The *tunnel valleys*[4] of Denmark and the *Rinnentäler*[5] of the North German Plain are said to have originated in this way. If this interpretation of their origin is correct, then by no means are all subglacial stream courses marked by eskers. According to this view esker streams are fully loaded and depositing whereas trench-cutting streams are underloaded and eroding.

Any turbulent stream, whether glacial or not, that drives sand or silt against firm bedrock, smooths the rock or fashions its surface into an irregular network of coalescent shallow cup-shaped depressions separated by narrow cusplike salients. Surfaces of this kind appear in many places beneath thin stratified drift in natural or artificial exposures and

[1] Good general references include Gilbert 1906; Tarr and Martin 1914; von Engeln 1911.
[2] Gilbert 1906, p. 317; Riley 1943.
[3] Hudson 1912.
[4] Madsen and others 1928, p. 108.
[5] Woldstedt 1929, pp. 75–85.

testify to the eroding action of the stream that deposited the drift. At first sight the smooth, polished rock surfaces resemble glacially polished rock, but on further examination they are seen to differ in that they lack striations and are smoother in the hollows than in the intervening salients, whereas with glacial polish the reverse is usually true.

Ice-Marginal Streams[6]

Spur Notches and Other Short Superposed Channels

The erosional effects of ice-marginal streams are both numerous and striking, both in regions of hills and mountains and in plains country. Probably these effects would never have been fully understood if similar features had not been observed in process of formation along the margins of still-existing glaciers. No better description of these features exists than the one written by Tarr after a visit to coastal Alaska:

> Along the margin of every glacier that reached well down into the zone of melting, in the Yakutat Bay region, there is a marginal channel, with ice for one wall and the mountain side for the other. These marginal channels are noted for their lack of continuity. In some places, after following the margin of the ice for awhile, and trimming the mountain slope, they disappear beneath the ice; or they may flow on a gravel bed which rests on buried ice; or they may leave the ice margin entirely and cut a gorge across a rock spur, perhaps returning to the ice margin lower down, or even going off to the sea by an independent course. Where engaged in gorge cutting these streams work with great rapidity, for the volume is great, the sediment load heavy, and therefore, with sufficient grade to prevent deposit, they are active agents of erosion.
>
> When such channels are found in process of formation their characteristics are easily observed; but above these, marking former higher levels of the ice, are others now abandoned. These are far less easy to understand except in the light of those below, now forming along the existing ice margin. There may be a short section of gorge, contouring the hillslope and open to the air at both ends; or the gorge may extend, on its lower end, across a divide into another valley, and in its lower course appear to be a normal stream gorge though its upper course is most unnatural without taking into account the former presence of the ice. There may be steep cliffs, evidently trimmed by stream erosion, whose formation is difficult to understand unless one can restore the old ice wall that once formed the other bank of a stream valley. Sometimes these trimmed cliffs end in a gorge like the above. Sometimes they die out without any apparent cause; but this

[6] See von Engeln 1911; Waters 1933; Kendall 1902.

> is easily understood when one pictures the stream disappearing under the ice, or flowing on it. At times there is a perfect stream bed at the base of the trimmed cliff; but at other times there is only a narrow terrace, often of very irregular form. Then it requires much imagination to restore a stream valley here where we must supply not only one wall but even the stream bed. The vanished glacier readily accounts for it, however; and if one doubts the explanation, he has but to go a few hundred feet downward to the place where a stream is actually flowing on ice veneered with gravel, and walled by a moraine-covered ice bank on one side and a trimmed cliff on the other. Here it is all clear, for all the elements are there; but in the channel above three of the four elements of the ancient valley are gone — one wall, the stream bed, and the stream itself.[7]

Both Tarr and von Engeln have shown that marginal drainage is more vigorous during the shrinkage of glaciers than during their expansion. In fact most of the sharply defined features cut by marginal streams, and clearly visible today, were made during the last great deglaciation.

The valley-side trenches cut by marginal streams range from little trenches only a few feet deep[8] to great gorges hundreds of feet in depth.[9] Where a glacier occupies a valley having closely spaced tributaries, the marginal stream becomes superposed across a series of spurs, cutting into them notches aligned along the stream course and hence more or less parallel with the trend of the main valley.

Some of the gorge-cutting streams were the outlets of lakes held against slopes by a wall of glacier ice. Chains of lakes and connecting streams are clearly recorded in the Cleveland Hills in Yorkshire, England, and have been fully described in a classic study by Kendall.[10]

Rates of cutting by marginal streams vary widely but are likely to be high because of steep gradients, ample loads, and frequent though short-lived flood discharges. A small stream from Casement Glacier, Glacier Bay, Alaska, was superposed across resistant bedrock in 1935. When seen in 1941 it had cut a gorge 50 feet wide and more than 25 feet deep.[11] Because of the rapid and variable rate of gorge cutting, estimates of time based on this factor are likely to be very unreliable.

Gorges and channels cut by streams superposed from ice-marginal positions must be carefully distinguished from channels cut by proglacial or postglacial streams superposed from outwash, till, or other thick glacial

[7] Tarr 1909*a*, p. 101.
[8] Mikkola 1932, p. 47; Mannerfelt 1945, pp. 16–21.
[9] Flint 1935*a*, pp. 173–177.
[10] Kendall 1902.
[11] Wm. Osgood Field, Jr., *unpublished*. For similar occurrences see von Engeln 1911.

sediments. After most of a sedimentary fill has been washed away the stream course superposed from it is easily confused with an ice-marginal course, yet the district in which it lies may never have been glaciated.

Large Ice-Marginal Rivers

Abandoned segments of large river valleys, some of them also filled with drift, show that at times an ice sheet filled part of a valley, ponded the drainage, and forced it to spill out and cut a new course detouring the ice sheet. No two cases are exactly alike because of variations in the relation of the ice margin to the general slope of the land and to other pre-existing valleys.

A conspicuous example is the temporary diversion of the Columbia River through the Grand Coulee in the State of Washington. During some glaciation that preceded the latest glaciation of northern Washington the Okanogan lobe of the Cordilleran ice blocked the east-west canyon of Columbia River and more than filled its maximum depth of about 1500 feet. The river formed a lake above the dam and rose until it spilled over a local threshold into the pre-existing valley known as Grand Coulee. The great discharge that passed through the coulee formed several conspicuous falls,[12] and following existing low-level routes detoured the ice and spilled back into the Columbia at several points. The farthest of these was nearly 100 miles from the point of detour. Other, shorter detours were forced upon the Columbia west of the Grand Coulee.[13] During a later glaciation the river was again blocked at the same place, and there was formed a glacial lake that also escaped via the Grand Coulee route.[14]

Farther east, the middle and upper segments of the Missouri River underwent great changes because of blocking by the ice sheet.[15] The upper and lower segments of the river flow eastward down the regional slope, but the middle segment, in the Dakotas, virtually contours the regional slope through a distance of nearly 300 miles. It has been generally inferred that much of this segment came into existence during one of the earlier glaciations. It closely parallels the border of the drift (Fig. 43) but clearly antedates at least one ice invasion. Abandoned valleys partly filled with drift lead eastward from the eastern side of the Missouri River trench, in apparent continuation of valleys entering the Missouri from the west.

Along the upper segment of the Missouri in eastern Montana similar

[12] Bretz 1932.
[13] Flint 1935*a*, p. 176.
[14] Flint and Irwin 1939.
[15] See Todd 1914; Leonard 1916.

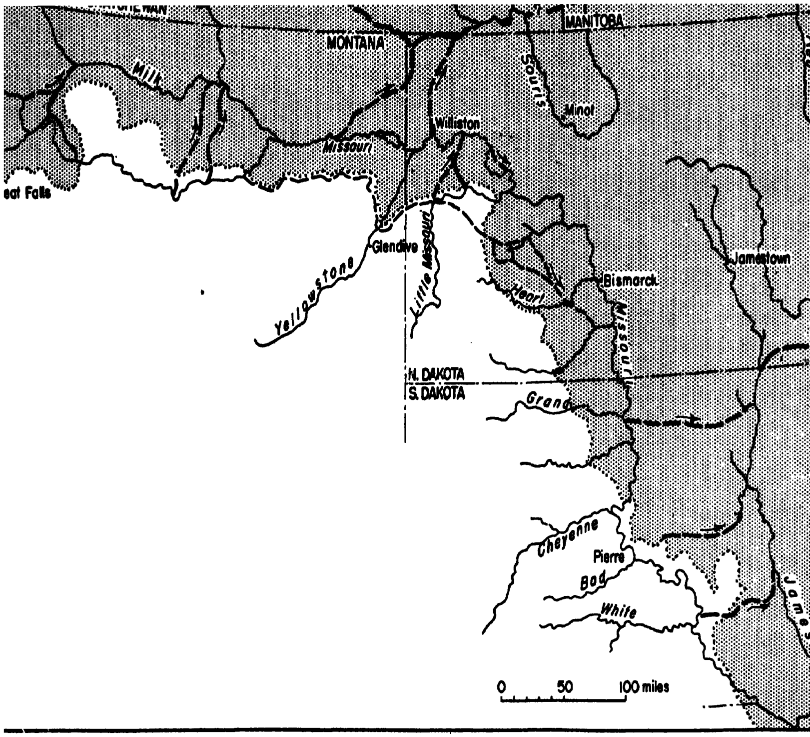


FIG. 43. Present and former drainage lines in the region of the middle and upper Missouri River. (Compiled from Alden 1932, Leonard 1916, Todd 1914, and other sources.)
Thick broken lines = inferred preglacial stream routes.
Thin broken lines = inferred temporary, ice-marginal stream routes.
Shaded area = glaciated region (boundary approximate only).

abandoned and partly buried valleys lead northward to the Milk River, and other abandoned valley segments suggest that prior to glaciation the Missouri, joined by the Yellowstone and the Little Missouri, flowed north into the Souris-Assiniboine system, perhaps discharging via the Lake Winnipeg depression, the Nelson, and the Hudson Bay region. The advent of the ice sheet flowing from the northeast blocked all this drainage and detoured the Missouri along the ice margin. That this event happened more than once is indicated by an abandoned channel system more than 300 miles long, extending from the upper Missouri in eastern Montana to the middle Missouri south of Bismarck in North Dakota. Throughout nearly half this distance the channel coincides with the margin of the glaciated region. Channel cutting throughout this region was comparatively easy because the bedrocks are generally weak, with shales predominating. The glacial geology of this region is so little known that only the general outline of its drainage history can be laid down at present.

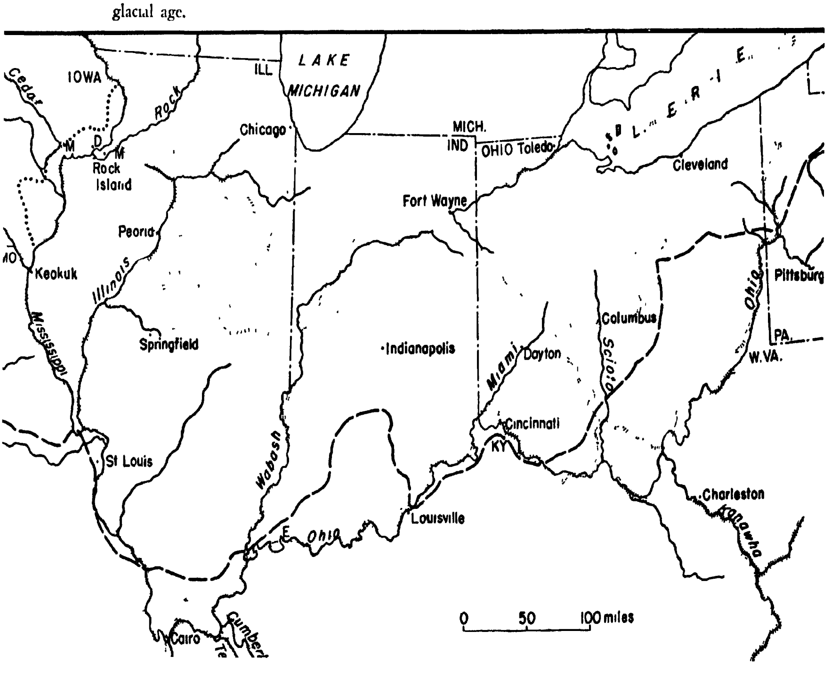
Farther south, at Kansas City, the Missouri River emerges from the glaciated region. Here its trench abruptly widens and turns sharply eastward. At the same point the large Kansas River enters from the west. These relations are interpreted to mean that the Missouri River below Kansas City is the preglacial continuation of the Kansas River, and that the segment above Kansas City came into existence much later as an ice-marginal stream.[16]

A less spectacular but more complex case is that of the Mississippi River in Iowa and Illinois. Here the detour segments have been in part filled with drift during subsequent glacial ages and still later in part re-excavated. From the drift fillings and the vertical relations of the bedrock floors of the various valley segments the general sequence of events has been provisionally worked out.[17] The preglacial courses of the Mississippi and some of its tributaries are shown in Fig. 44. In the Nebraskan or Kansan age the ice sheet forced two detours, one of which (between Rock Island and Muscatine) is followed by the river today.

[16] Greene and Trowbridge 1935.
[17] Leverett 1921; 1942.

FIG. 44. Drainage changes in the Ohio and Mississippi river basins caused by successive glaciations. (Compiled from Stout, Ver Steeg, and Lamb 1943; Fidlar 1943; Horberg 1945; Leverett 1942; and other sources)

Double fine dotted lines = conspicuous preglacial streams.
Thick broken line = outer limit of glacial drift.
Coarse dotted line = temporary ice-marginal route of Mississippi River during Illinoian glacial age.



In the Illinoian age the ice sheet flowed in from the east and forced a temporary detour from Savanna to Keokuk. During the Wisconsin age the extent of glacier ice in this region was insufficient to produce any notable change in the river's course.

Far more extensive changes were accomplished by the ice sheets in the drainage of the Ohio region. Compilation of thousands of well records has aided in the recognition of a system of valleys filled or partly filled with drift related to one of the earlier glaciations. From this the main outlines of the preglacial drainage have been reconstructed[18] (Fig. 44). The Ohio River then extended little east of Cincinnati, and what is now the upper Ohio flowed north, probably to a river flowing northeast through the Lake Erie basin. Larger than either of these streams was the Teays[19] River, a major prolongation of the Kanawha River. The Teays flowed northwest across Ohio and Indiana, and apparently also across Illinois, occupying a wide, deep trench comparable in size with the trench of the lower Ohio River today.

With the advent of the ice sheet during the Nebraskan Glacial age the lower Teays basin was invaded by glacier ice and long finger lakes were created by the water in the Teays system dammed by the ice in Ohio and West Virginia. Thick sediments, accumulated in these lakes, remain in some of the valleys. The evidence indicates, not the cutting of new channels along the ice margin, but rather the integration of east-west drainage lines at some distance beyond the edge of the glacier. This seems to have been accomplished by the trenching of two divides that separated the Teays system from its neighbors. In this way the present-day Ohio River, far longer than its preglacial predecessor, came into existence. The Kansan ice filled the old valley system with drift so that it ceased to function as a valley.

In northern Europe the drainage was radically altered by glaciation. At its greatest extent during the Fourth Glacial age (and probably during earlier glacial ages as well) the Scandinavian Ice Sheet blocked the normal north-flowing drainage of the North German Plain. The lowest land lay along the margin of the ice sheet, descending toward the north and west. This route was followed by meltwater augmented by the discharge of the Weichsel, Oder, Elbe, Rhine, and Meuse rivers entering from the south. The floor of the southern part of the North Sea was then

[18] The compilation by Stout, Ver Steeg, and Lamb (1943) summarizes the extensive literature on this subject, with special reference to Ohio. See, for Kentucky, Leverett 1929*b;* for Indiana, Fidlar 1943; for Illinois, Horberg 1945.

[19] Named for an abandoned segment of its former valley in western West Virginia. The buried valley in Illinois has been called *Mahomet Valley.*

dry.[20] The combined ice-marginal river followed this toward the southwest, into the Strait of Dover (also dry) and the English Channel. Joined by the Thames, Seine, and other rivers plus meltwater from the British glaciers this "Channel River" entered the sea (which then stood perhaps 150 to 200 feet lower than now) between Brittany and Cornwall. The part of the route that is not now submerged is marked by stream trenches and sediments that form a system described below.

Stream-Trench Systems

The route of the "Channel River" is characterized not by a single continuous trench but by a system of subparallel interconnecting trenches, partly filled with outwash and other sediments, that form a vast complex network. As the Scandinavian Ice Sheet shrank from northern Germany and Poland, its retreating margin continually uncovered new and lower ground and thus opened new and lower routes for the ice-marginal drainage that was escaping westward. Routes were successively abandoned in favor of parallel routes farther north. The diversions took place partly through uncovered segments of preglacial north-draining valleys and partly along *Rinnentäler,* believed to have been excavated by subglacial streams, and exposed by deglaciation. The resulting complex of interconnecting trenches is striking (Plate 4). German geologists call them *Urstromtäler,*[21] and, because they were formed at various times and are complicated by refillings with drift, including extensive ice-contact deposits, and partial re-excavations by the ice-marginal streams, their detailed history has not yet been fully worked out.

In southwestern Russia the area glaciated by the Dnepr and Don lobes of the Scandinavian Ice Sheet during the Third Glacial age includes a similar though less extensive system of stream trenches, most of which were cut along lines determined by the temporary position of the glacier margin.[22]

Another similar though less extensive system of drainage trenches, many of them marginal, is cut into the plains of southern Alberta.[23] These, like the European systems mentioned, are in a region of low relief and comparatively weak rocks, which, being easily eroded, permit the

[20] Configuration of the North Sea floor is well shown on a map by R. G. Lewis (1935). There is general agreement as to the existence of the meltwater stream, but difference of opinion as to its exact course through the region now submerged.

[21] For a good résumé see Woldstedt 1929, p. 171; also Lencewicz 1927.

[22] Spreitzer 1941, p. 15.

[23] Bretz 1943; M. Y. Williams 1929.

cutting of a large number of deep trenches within a comparatively short time.

The chief value of ice-marginal streamways is that they afford a means, in addition to end moraines and outwash heads, of determining the successive positions occupied by glacier termini during a time of general deglaciation.

Ice-Marginal Lakes

Lakes formed by blocking of normal drainage by glacier ice are rare today because most existing glaciers lie on high land that drains away from the ice. However, during the times when the former great ice sheets were extensive many lakes, some of them very large, were formed in this way. Most of them date from the latest glacial age.

Nevertheless in favorable places we find remnants of lake sediments dating from earlier glacial ages, and it is probable that in those times lakes were nearly as numerous and extensive as during the last great deglaciation. Their sediments have been destroyed by erosion during the interglacial ages or concealed beneath the drifts of later glaciations. The best-preserved records of early glacial lakes in North America are the extensive Minford silts in southeastern Ohio[24] formed by glacial blocking of north-flowing streams in Nebraskan or Kansan time, before the long Ohio River of today had come into existence.

The stratigraphic positions of most of the extensive bodies of lake sediments show that the lakes came into existence during the shrinkage rather than during the expansion of the last ice sheets. No doubt water was ponded during the preceding phase of glacier growth, but time was shorter and production of meltwater was less than during the phase of shrinkage. Then the expanding glaciers overrode, reworked, and covered with till the deposits previously made beyond their margins. In summary, the late date of the existing record of ice-marginal lakes is not an indication either that lakes were formed only during deglaciation or that the latest glaciation differed in any essential respect from the earlier ones. Rather it is a function of degree of preservation of the record.

The great lakes that formed around the fringes of the shrinking Laurentide, Scandinavian, and Siberian ice sheets are briefly described in Chapters 13, 15, and 17 respectively. The areas of the former North American lakes are shown on the Glacial Map of North America.[25]

Proglacial Streams

The valleys that led away from the former ice sheets discharged great volumes of meltwater. That their loads of sediment were in general

[24] Stout and Schaaf 1930.
[25] Flint and others 1945.

much greater than their present nonglacial loads is shown by the bulky sedimentary fills deposited in these valleys by the meltwater streams; the modern streams are removing these fills. Large streams such as the Columbia and the Susquehanna, whose lower courses extend beyond the glaciated country, as well as streams draining the territory progressively uncovered by the glaciers, were profoundly altered by the temporary but great discharge of water and sediment.

Proglacial Mississippi River System

The greatest proglacial drainage system in North America was the Mississippi-Missouri-Ohio system, which drew its discharge from a sector of the ice sheet nearly 2000 miles long. Even at the times when the ice sheet was at its maximum extent the combined discharge of these rivers probably was greater than it is today, because precipitation must have been greater. During the times of most rapid glacier shrinkage, when stored-up frozen precipitation was being melted and discharged much faster than it was being replaced, the discharge was surely very much greater than at present.

The proglacial regimens of these streams differed from their present condition. Instead of being relatively deep and confined in most places to a single channel, each stream consisted of a braided network of shallow channels. This is inferred from the form and stratification of the fill built up by each river within its trench. In the Mississippi trench the fill built during the Wisconsin or Illinoian Glacial age is the only one whose remnants are extensively exposed to view. In northern Illinois the upper surface of the Wisconsin fill stands 100 feet above the present river, which in turn is flowing on deposits that in places reach more than 100 feet above the bedrock floor of the trench. As far downstream as northern Arkansas, outwash is 160 feet thick and its top stands 60 feet above the stream.

Proglacial Drainage on the Columbia Plateau

Probably the most remarkable effects of glacial drainage yet described occur on the Columbia Plateau in eastern Washington, where a thick succession of basalts are warped to form a broad east-west syncline, the Pasco basin. The northern part of the area was widely and repeatedly overrun by glacier ice. Meltwater flowed southward down the dip slope, eroding the basalt to form complexes of hills, basins, and channels, widening preglacial valleys, and depositing great volumes of basalt debris mixed with sediments contributed by the glacier. The remarkable complex of valleys and channels is widely known as *scabland,* and the combined volume of erosion and deposition they record is very great.

The only aspect of the problem on which agreement is complete is that the scabland is the work of glacial meltwater. Beyond that point opinions differ, and three principal hypotheses have been advanced in explanation of how the work was accomplished. Bretz ascribed the scabland to the operation of a catastrophic, short-lived flood of prodigious volume which enormously enlarged the pre-existing valleys and built within them great constructional "bars."[26] Allison ascribed the "bars" and many of the erosional features of the scabland to streams diverted along and around the sides of major valleys that were blockaded by prodigious accumulations of berg and river ice piled up in vast jams.[27] Flint regarded the "bars" as remnants of far larger masses of sediment that formed thick continuous fills aggraded by normal proglacial streams to the level of a temporary lake in the Pasco basin. The erosional features in the bedrock were attributed by him to the repeated superposition of streams as they dissected the fills.[28]

The sediments involved are unusual in that they consist predominantly of material picked up by the streams after they had left the glacier margin. Sediments of glacial origin, however, occur down the Columbia almost to its mouth, more than 300 miles beyond the glacial source of the meltwater.

The Proglacial Volga

The greatest of the European proglacial streams (as distinct from ice-marginal streams) was the Volga. During the Third Glacial age it drained a sector of the ice margin nearly a thousand miles long, reaching from the Ural Mountains to the region of Stalingrad. During the Fourth Glacial age the glacial sector drained by the Volga was almost as long, but the Volga system was much longer than in the earlier time, because the ice sheet did not reach as far southeast. During the glacial ages the Volga was considerably shorter than now, because the greatly increased Caspian "Sea" then submerged the lower course of the river through a distance that at one time, at least, exceeded 600 miles. It was as though the Mississippi Valley had been submerged as far north as St. Louis.

Summary

From the examples cited it is apparent that meltwater streams greatly extended the indirect influence of glaciation over a very large territory that was not itself glaciated. The outwash fills of these streams constitute a means of correlation of the greatest potential value between the drift

[26] Bretz 1928 and other papers.
[27] Allison 1933; 1941.

sheets on the one hand and the fluctuating sealevel on the other. Hitherto the fills have been given much less study than they deserve, but the results of such studies in the future will go far to establish relationships between glaciation and events in the extraglacial regions.

ICE-THRUST FEATURES[29]

Deformation of stratified drift and even of nonglacial material[30] by thrust as the surface is overridden by an active glacier is shown both by topography and by vertical sections of the overridden deposits. The topography commonly consists of low irregular ridges, smooth or undulatory. These are the *Stauchmoränen*[31] of Gripp, who ascribed this origin to many moraines in northern Germany. The irregular thrust topography was referred to by Woodworth and Wigglesworth as "pseudomorainal."[32]

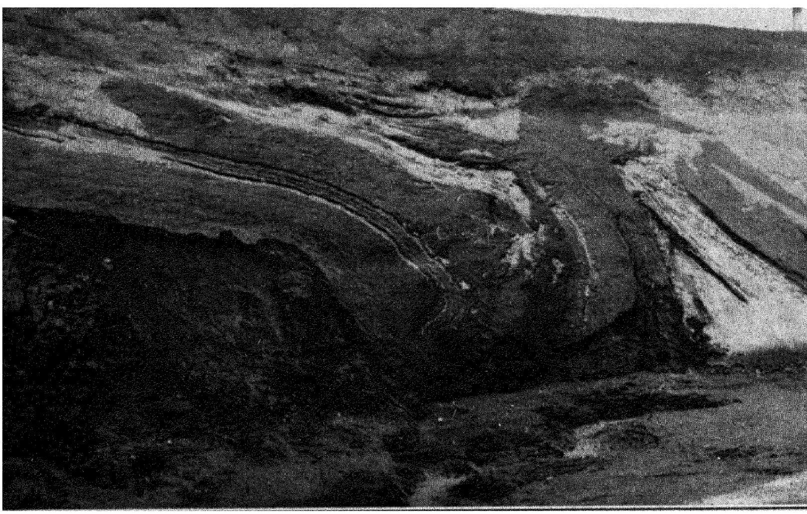
Vertical sections in some such overridden areas reveal folding and irregular contortion, with the fold axes transverse to the direction of thrust of the glacier, and some of them arcuate in plan. The most common folds are small sharp asymmetric flexures (Fig. 45) that may be isoclinal, overturned, or associated with thrust faults. Such folds are of small extent and die out downward, rarely reaching 50 feet in amplitude; they range down through intermediate stages to broad gentle flexures. On Long Island, New York, flexures of this kind are 50 miles long, several miles wide, and 50 to 100 feet in amplitude.[33] They involve beds of clay, whose compaction under the weight of the edge of the ice sheet, and consequent flow, may have been largely responsible for the flexures. Such an origin implies that the subglacial materials, being free to flow, were not frozen. The sharp minor folds, on the other hand, apparently could have been made whether the sedimentary layers were frozen or not, provided only they were subjected to sufficient confined pressure.

In some places overriding results in erosion rather than contortion and ridging. The beveled top of the fold shown in Fig. 45 suggests that planation followed the folding. In coastal Alaska Tarr found thick outwash gravel that had been overridden by a glacier. The upper surface of the gravel was irregularly undulatory, the undulations truncating the stratification of the outwash. He inferred that the gravel was frozen solid at the time of overriding; otherwise the stratification would have been dis-

29 See M. L. Fuller 1914; Sardeson 1906.
30 Carney 1907.
31 Gripp 1938.
32 Woodworth and Wigglesworth 1934, p. 67.
33 M. L. Fuller 1914, p. 203. See also Hopkins 1923.

turbed and the component stones eroded piecemeal. As Tarr observed, erosion of any frozen mass must take place by abrasion, for there would be no joints to permit quarrying.[34]

Very commonly clay-rich till, apparently deposited by lodgment, overlies stratified drift without either disturbance of stratification or visible erosion. This might be a result of a very considerable load of drift in the basal part of the overriding ice. Erosion would thereby be prevented and deposition would take place instead. Truncated folds overlain by till might indicate deformation and erosion by relatively drift-free ice, followed by deposition as the glacier acquired a basal load.



Henry Retzek

Fig. 45. Detail of fold in stratified silt and sand, made by thrust of overriding glacier. Near Fergus Falls, Minnesota.

Some of the eskers in Denmark contain at the center of the base a low ridge of clay-rich till paralleling the trend of the esker. Andersen believed the ridge to have been squeezed up into the esker tunnel, before esker deposition, by the static pressure of the ice upon its plastic till floor on both sides of the tunnel.[35]

The most penetrating inquiry into the phenomena of ice thrust has been that of Slater.[36] He classified the structures into two distinct types. The first is caused by drag of glacier ice over weak underlying strata, producing displacement and folding somewhat analogous to the wrin-

[34] Tarr 1909*c*, p. 126.
[35] Andersen 1931*a*, p. 178.
[36] Slater 1926; 1926–1927*a*; 1926–1927*b*.

kling of a tablecloth by a hand sliding across the table. The second type is the structure acquired by the drift while actually incorporated in moving glacier ice and preserved throughout the period of slow wastage of the ice. This structure has been termed "glacial-pseudomorph structure." It is characterized by imbricate or *Schuppen* structure and is believed to represent the upthrusting observed in existing glaciers and occurring immediately upstream from the stagnant terminal part of the ice. This terminal ice creates an obstacle which the flowing ice upstream attempts to surmount. Thrusting is the normal result.

The first type of structure described above is regarded as occurring first in time, commonly during glacier expansion, when the ice is thick and vigorous. The resulting tectonic features (which also constitute ridges and other topographic features) act as a nucleus over which structures of the second type are molded in the terminal zone of the glacier. The latter structures are built in the terminal zone during glacier shrinkage when the ice has become thinner and is flowing sluggishly. The process is likened in some degree to the process of accretion of till around a nucleus during the construction of a drumlin.

There is little doubt that insufficient attention has thus far been paid to the structures produced in till and other materials by the movement of glacier ice. Further research in this field is likely to result in clarification of the problems involved in the movement of glaciers in their terminal regions.

EOLIAN DEPOSITS

Sand Dunes[37]

Position, Form, and Stratification

Sand dunes built wholly or partly of reworked glacial drift occur in many places. Some have ceased to accumulate and have become covered with vegetation. Others are in process of construction. But all are in the immediate proximity of the glacial deposits that are their source of supply.

Most of the dunes occur in groups, clusters, or "fields." The majority of them rest upon or at the margins of conspicuous masses of outwash. In northern Germany, for example, many large dune groups occur in the *Urströmtäler*. This, together with their sedimentary content, indicates that they consist largely of redeposited outwash.[38]

Other dune groups are linear and lie close to strandlines (usually

[37] For comprehensive European studies see Högbom 1923; Cailleux 1942. A good specific American study is Cooper 1935.

[38] The grain-size relation of dune sand to the outwash from which it was derived is discussed by Cooper (1935, p. 99).

beaches) of glacial lakes, or the sea, formed mainly during the last great deglaciation and in many cases subsequently removed from the water's edge by crustal warping or lowering of the water level or both. These dunes, less abundant than those built directly from outwash, consist of sand that has been twice reworked, first by waves and later by the wind.

Still other groups of dunes are related to the shores of pluvial lakes in nonglaciated regions — lakes that have since shrunk or dried up entirely. Many of the dune groups in the Basin-and-Range region of western United States were built of sand from such sources.

On a basis of their form three general types of dunes are distinguished. (1) *Bow-shaped wavelike dunes* are characteristic of most large dune areas. In ground plan they are convex in the direction toward which the wind was blowing. They lap over on each other like waves in a choppy sea. Individual groups of such dunes may be many square miles in area. (2) *Linear dunes* are common along strandlines, at the crests of scarps in terraced outwash, and at the bases of hills projecting through outwash masses. Such dunes are built close to the source of sand, wherever obstacles for lodgment of the sand are offered, regardless of prevailing wind direction. (3) *Parabolic dunes,* commonly wishbone-shaped in ground plan, are convex in the direction in which the wind was blowing. These are probably the least common of the three types, although they are numerous on many parts of the Great Plains in Canada.

The bearing of the form and stratification of dunes on the climates and wind directions that prevailed when they were built is discussed in Chapter 20. In general, within the belt of westerly winds, dunes tend to be concentrated at the eastern (leeward) margins of outwash bodies and on west-facing shorelines and terrace scarps. However, as they are not invariably thus concentrated, dune positions are not a sure basis for inferring direction of the wind at the time of accumulation.

Time and Conditions of Origin

The range of conditions under which dunes can form is so large, as demonstrated by the facts of the dunes themselves, that whether any general inference can be drawn as to an optimum time of dune building is doubtful. It is stated in Chapter 20 that neither lack of vegetation nor arid climate is required for dune accumulation. A sufficiently dry sand surface is all that seems to be necessary. This condition, depending to a considerable extent on fluctuation of the ground-water table, could have been, and probably was, reached at very different times, relative to outwash deposition, in different places. In consequence some dune groups were accumulated while outwash deposition was actively in progress

whereas others did not come into existence until dissection of the outwash had begun. This statement is not meant to imply that widespread climatic desiccation such as occurred some 6000 years ago (Chapter 21) did not result in dune-building activity. It implies merely that climatic change is not a necessary prelude to dune construction and that many dune groups have come into existence through quite other causes.

In north-central Europe and elsewhere the dunes and the loess have been held to be contemporaneous, the dunes constituting a northern belt of coarser sediment dropped first, and the loess a more southerly belt of finer sediment carried farther, by northerly winds blowing from the ice sheet. The implied gradation of grain size, however, has not been established. Further, Högbom believed that the main body of loess in Europe is much older than the dunes of northern Europe, having been deposited at the maximum of the Fourth Glacial age. The dunes he views as much later, having been built after westerly winds had wholly replaced the northerly winds that blew from the Scandinavian Ice Sheet when it was extensive.[39]

Loess[40]

Character and General Origin

One of the most remarkable of the Pleistocene deposits, around which an extensive literature accompanied by much controversy has accumulated, is the *loess*. Because of former widely differing views as to its origin, this material has been given, specifically but more often by implication, many different definitions. However, most of them are unsatisfactory in one way or another. Hence, at the risk of adding to the already burdened literature on the subject, it seems best to frame a definition if only as a basis for the present discussion.

Loess, then, is a buff-colored nonindurated sedimentary deposit consisting predominantly of particles of silt size. Commonly it is nonstratified,[41] homogeneous, calcareous, and porous, and it may possess a weak vertical structure resembling jointing.

Only a part of the loess can be properly classed as of glacial origin, but even this part is abundant and in most places stands in a fairly definite relation, both areally and stratigraphically, to deposits of unequivocal drift. Furthermore, as it occurs in broad continuous sheets, it is an important aid in identifying the sheets of till with which it is interbedded. For

[39] Högbom 1923, p. 232.

[40] Good general references are Scheidig 1934; Grahmann 1932; Stuntz and Free 1911.

[41] Very limited pockets of stratified silt occur in the loess. These are widely interpreted as a local pond-deposited facies of an otherwise eolian sediment.

these reasons loess is a major element in the Pleistocene deposits in and near the glaciated regions of the world.

Mechanically loess consists in large part of silt (grains 1/16 to 1/256 mm. in diameter) with subordinate clay and traces of fine sand (1/8 to 1/16 mm.). Mineralogically it is made up principally of quartz, with smaller amounts of clay minerals, feldspars, micas, hornblende, and pyroxene. Carbonate minerals are variable, ranging as high as 40 per cent. This composition is so elastic that it tells little about the rocks in which the minerals originated.

Where not weathered, loess is gray; but, as permeability facilitates deep oxidation, the material appears in most exposures as buff, tan, or reddish. For the most part it lacks stratification, but local lenses and pockets in loess may be well stratified, usually with delicate lamination. Fossils, which are abundant in some parts of the loess, consist largely of woodland snails plus a few aquatic snails. In addition there are a few land mammals, mainly rodents but including musk-ox, bison and mammoth.

The sheets of loess are in most places only a few feet or a few tens of feet in thickness. Locally along the Mississippi River, however, loess is 100 feet thick. In Nebraska thicknesses up to 168 feet have been measured, and in Eurasia even greater thicknesses are reported. The thin sheets cover hill and valley, having a great vertical range (1000 feet in the Mississippi basin) and being consistently thicker on the leeward sides of hills and bluffs than on their windward sides.

By almost all geologists today the bulk of the loess is regarded as having been transported and deposited by the wind.[42] The two chief sources of fine sediment exposed at the surface without the protection of a continuous cover of vegetation are desert basins and active outwash plains. The distribution of loess is so closely related to the distribution of such sources that there is little doubt about the general source areas of loess.

This fine sediment is widespread to leeward of many desert basins, particularly those in central Asia, and forms thick coatings over the mountain slopes immediately adjacent to the basins. Aside from the outwash from comparatively small areas of glacier ice on the mountains surrounding the basins, most of this loess is of nonglacial derivation.

Where the loess of glaciated regions has been closely studied it has been found that it becomes finer-textured, and also thinner, with increasing distance from its source in the drift. It has been found also that the mineral content of the loess resembles that of the corresponding size-

[42] This is clearly brought out by the summaries in Scheidig 1934, Grahmann 1932, Obruchev 1945, W. B. Wright 1937, Stuntz and Free 1911, and Thwaites 1941.

grade fraction of the till in the same region.[43] According to Grahmann[44] loess derived from deserts can be distinguished from loess derived from outwash by means of the range of its grain size. Loess derived from deserts has a much wider range of sizes, including a conspicuous quantity of very fine grains. This comparatively poor sorting reflects the fact that the sediment has been sorted only once, during its transport by the wind. Loess derived from outwash, in contrast, has but a narrow range of grain size, the coarser and finer particles having been screened out. This more thorough screening reflects a double sorting, first by the meltwater streams and then by the wind.

Loess and loesslike sediments may also have originated in favorable localities through weathering and mass-wasting processes,[45] and eolian loess may have been reworked by mass-wasting. It is improbable, however, that any considerable proportion of the loess in and near the glaciated regions was finally deposited in this manner.

Because loess is confined almost entirely to the Pleistocene strata, it has sometimes been thought to record a unique (and in some opinions a mysterious) set of climatic conditions peculiar to an "ice age." But with more detailed study the loess has gradually lost whatever mystery may have attached to it. It is true that the Pleistocene, with its high land barriers and its desert basins in their lee, and with its extensive areas of rapidly accumulating outwash, afforded opportunity for the deflation of silt and clay on a scale far greater than that which has obtained with the low lands and maritime climates of the greater part of geologic time. On a reduced scale, however, winds have doubtless made extensive deposits of silt throughout geologic history. Pre-Pleistocene rocks believed to be consolidated loess have been reported.[46] But in general the older loesses, even the older Pleistocene loesses, have been removed by erosion or have lost their identity through mixing with other kinds of sediment. The Pleistocene epoch provided conditions ideal for loess making, but it was not uniquely the workshop for eolian activity that it was once thought to be.

With this general statement we may turn to the loess of the glaciated regions, for it is these with which we are chiefly concerned, although (if we may disclaim any intention to pun) desert winds have given rise to large volumes of loess. The glaciated regions in which the loess has been most thoroughly studied are central North America and central Europe.

[43] Kay and Graham 1943, p. 183; see also Savage 1915.
[44] Grahmann 1932.
[45] Cf. Berg 1932; R. J. Russell 1944.
[46] Cf. Obruchev 1945, p. 257.

Loess in North America

AREAL DISTRIBUTION. Figure 46 shows the distribution of conspicuous loess in central North America. Two distinct loess regions are recognizable. The first lies in western Kansas and Nebraska. (Loess is present more or less continuously northward into Saskatchewan and Alberta, but it is so little known that it is not shown on the map.) This is a region with a steppe climate and a source of sediment to windward consisting of

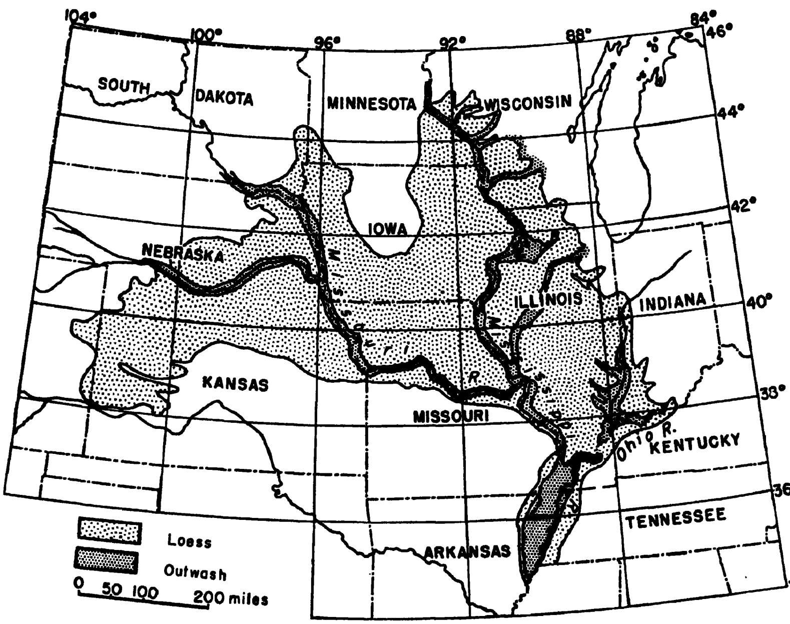


FIG. 46. Distribution of loess in central North America. (Generalized from E. T. Apfel *in* Flint and others 1945.)

extensive exposures of Cenozoic strata. In Nebraska, at least, the loess has been observed to grade into sand in the vast area of the Nebraska Sand Hills, and this in turn was derived from the bedrocks.[47] The loess in western Kansas and Nebraska, therefore, is mainly nonglacial. In that fact it resembles the bulk of the loess deposits in central Asia.

The second loess region is closely related to the Missouri and Mississippi rivers and some of their tributaries, notably the Ohio and the Wabash. Loess is conspicuously better developed on their eastern (at present leeward) sides than on their western sides. It reaches its maxi-

[47] Lugn 1935, p. 161; Bollen 1945, figs. 1, 2.

mum thickness, in places approaching 100 feet, on the crests of the eastern bluffs. Away from the bluffs it thins rapidly, thinning within a few miles to perhaps 10 or 15 feet. In some areas close to its outwash sources the loess forms a constructional topography, but throughout most of its extent it merely mantles pre-existing land forms.

The rivers mentioned above deposited huge volumes of fine outwash in the form of valley trains, whose terraced remnants are still strongly evident in the river trenches. Throughout the region the sheets of loess lie on drift, immediately beyond drift, or between two drift sheets.

From this group of facts it can be inferred that the loess in this second region is derived mainly from outwash. In eastern Nebraska and Kansas, however, it must be adulterated to an unknown though doubtless very large extent with sediment derived from the western, nonglacial region. In the region farther east also the loess must include some nonglacial sediment. Leverett, a lifetime student of the loess, believed that the bulk of it in central North America was derived from nonglacial sources.[48] A less extreme position was taken by Lugn, who nevertheless stressed nonglacial sources.[49] On the other hand Kay and Graham held the loess in Iowa to be of local origin.[50] With adequate mechanical and mineralogic analyses of the loesses and their probable source sediments this problem could be solved, but no general solution has yet been reached.

In both regions of central North America the loess seems to demand westerly and southwesterly winds, like the planetary winds prevailing at present, as the chief agents of transport. No doubt winds from other directions, including winds blowing from the region of high pressure over the ice sheet to the north, aided in distributing the loess, but they do not appear to have been the dominant factor.

Farther east loess occurs, but it is patchy in its distribution and has slight thickness. The loess in Indiana and Ohio is not thick enough to lend itself to ready mapping in the field. Still farther east it is only locally recognizable, as in the Boston region[51] and the Connecticut Valley.

At least three factors explain the scarcity or absence of loess toward the east. (1) From the Missouri River region to the Atlantic coast the climate changes from a steppe type to a moist continental type. The moister climate and the forest cover that accompanies it reduce surficial drying of the outwash in preparation for wind transport. (2) The drift in the Missouri River-Mississippi River region is rich in silt and clay, the chief components of loess, whereas the drift in the Appalachian region

[48] Leverett 1930*a*.
[49] Lugn 1935, p. 165.
[50] Kay and Graham 1943, p. 73.
[51] H. T. U. Smith and Fraser 1935.

and on the Atlantic slope is richer in sand and poorer in silt and clay. (3) The Missouri River-Mississippi River region has less relief and wider valleys than the Appalachian region and New England have; these conditions make for greater pickup of fine alluvium by the wind.

Thin loess overlying a Wisconsin drift occurs also in eastern Washington, a region of broad valleys and a dry steppe climate. In the same region the extensive "Palouse soil" formation consists of one or more earlier Pleistocene loess sheets derived from alluvium (not demonstrably outwash) to the southwest.

STRATIGRAPHIC POSITION. The stratigraphic position of the loess sheets is summarized in the following table because it has an important bearing on the conditions under which the loess originated:

Wisconsin Glacial stage	"Modern" loess	Very thin loess overlying the latest drift on the Great Plains and other parts of the United States and Canada.
	Tazewell Glacial substage	Thin loess overlying Tazewell drift in western Illinois.
	Iowan Glacial substage	Thick loess overlying Iowan drift and overlain by Tazewell drift. The surface loess of Iowa and of much of western Illinois; present in Nebraska.
Sangamon Interglacial stage		Loess overlying decomposed Illinoian drift and overlain by Iowan drift. Thick in southern Illinois; thinner in Iowa and Nebraska, where it is probably represented by much or all of the Loveland loess.
Yarmouth Interglacial stage		Loess overlying decomposed Kansan drift and overlain by Illinoian till, in Illinois. Loess outside the glaciated region in Nebraska.
Aftonian Interglacial stage		Loess overlying decomposed Nebraskan drift and overlain by Kansan drift in Iowa and Illinois.

From this table it is clear that loess is present in places on all the major sheets of glacial drift. The loess sheets of Iowan and Sangamon age are extensive; the extent of the older loesses is not known because they are seen in relatively few exposures. There can be no doubt, however, that these older loesses were originally more extensive than is evident from their present known distribution. They must have been largely reworked and removed by erosion, and perhaps also they are widely concealed from view beneath younger deposits.

The table shows a significant difference between the Wisconsin

loesses and the older loesses: the former are classed as glacial, and the latter as interglacial. This difference arises from the fact that the older loesses are commonly seen overlying deeply decomposed till, showing that they were not deposited until after long-continued interglacial weathering had run its course. The Wisconsin loesses, on the other hand, rest on till that is fresh or very little decomposed, showing that they were deposited very soon after (or during) the process of deglaciation.

This does not necessarily imply that in Wisconsin time conditions have been radically different from earlier glacial times, for two reasons. First, the Wisconsin Glacial age has not yet been succeeded by a long interglacial age of weathering such as the interglacial ages that succeeded each of the earlier glaciations. Because of this fact the Wisconsin section of drifts and loess sheets can not fairly be compared with the pre-Wisconsin sections. Second, we do not know that loess was not deposited during the various pre-Wisconsin deglaciations and later (*a*) removed by erosion, (*b*) buried and thus lost to view, or (*c*) decomposed and incorporated into the zone of deep decomposition that characterizes the upper part of each of the pre-Wisconsin tills. These zones may well include former loess, no longer distinguishable from the underlying till because of the radical decay that has affected the uppermost few feet of each section.

A large amount of interesting deductive argument and speculation has been devoted to the probable optimum time of loess accumulation.[52] With good reason the recent tendency in both Europe and America has been to regard the optimum time as the time of maximum glaciation. The character and positions of the Wisconsin loess sheets in North America, and in Europe as well, fully support this view. Along the Illinois River the mineral content of the loess is so related to stratigraphic zones in the adjacent outwash as to show that it was accumulated during the upbuilding of the outwash.[53]

The fossils in both the North American loesses — land snails,[54] mammoth, musk-ox, and bison — and the European loesses — land snails, musk-ox, reindeer, snowshoe rabbit, lemming, marmot, and the like — point to cold glacial rather than warm interglacial conditions.

Further, the pre-Wisconsin loesses in North America are not inconsistent with the view that they were accumulated during the glacial ages because the facts about them allow us to think of them as deposits

[52] See, for example, Visher 1922; Woldstedt 1929, pp. 121–124.

[53] Leighton 1932.

[54] Regarded many years ago by Shimek as recording a warm, dry climate, these fossils are now thought of as indicating cool, moist conditions.

made just before the glacial maxima and preserved through having been very soon buried beneath till. The loesses of comparable age, deposited over the upper surfaces of those same till sheets, must then be thought of as destroyed by decomposition and other forms of erosion during the subsequent long interglacial times. But until the stratigraphy of the loesses has been more fully studied this question will remain in the realm of speculation.[55]

Probably there is little uniformity in the rate at which loess has accumulated; the rate may be expected to have varied from time to time and from place to place. For example, the Iowan loess is inferred to have accumulated slowly throughout most of its extent, except near the margin of the ice sheet, where variable grain size and sparse vertical distribution of fossil snails suggest rapid accumulation.[56]

Nor can a uniform ecologic or climatic environment be assigned to the accumulation of loess. Fossils in the Iowan loess suggest forest in some places and grassland in others, and in at least one locality progressive change from forest to grass is indicated.[57] A forest cover should have protected accumulating loess from erosion more effectively than would a mat of sod. In Illinois, mollusks in Wisconsin loess show a transition from cold types at the base to warm-temperate types at the top, as well as an absolute increase upward in the mollusk population.[58]

The "modern" loess, thinner and less continuous than the earlier Wisconsin loess sheets, is developed primarily in the Great Plains region. Some of it may be related to the Mankato Glacial sub-age. But, having a steppe climate, this region continued to afford opportunity for the blowing of silt by the wind even after active outwash deposition had ceased. In central and eastern North America, however, the building of outwash on an extensive scale ended, in the Cary Glacial sub-age, with the creation of the glacial Great Lakes, which acted as traps for the silt formerly washed out along valley floors. Further, when the annual floods of silt ceased, the valley trains became more stable and were quickly converted into grassy meadows, thereby ceasing to be a source

[55] Because the Sangamon loess is somewhat leached beneath fresh Iowan drift, at least several thousand years must have elapsed between the end of loess deposition and the Iowan substage. We may speculate, though we can do no more as yet, that this loess was contemporaneous with a glacial expansion very early in the Wisconsin age, slightly earlier than the Iowan and of lesser extent. This is a purely *ad hoc* suggestion, but it is wholly reasonable. Alternatively there is the hypothesis that the Sangamon loess is truly interglacial and that it is largely of nonglacial origin, having been blown from the dry Great Plains region on the west. Only a detailed provenance study of this loess, lacking as yet, can fully settle this question.

[56] Kay and Graham 1943, p. 167.

[57] Kay and Graham 1943, pp. 190–191.

[58] Baker 1936.

from which silt could be seized by the wind. In the region inside the belt of extensive glacial lakes the rocks, unlike those of the Great Plains and the Central Lowland, are of types not productive of a great volume of silt and, furthermore, had been largely swept bare of loose surficial material by the ice sheets. Finally the climate is moist and the streams characteristically consist of chains of lakes in which the small loads of sediment settle, safe from seizure by the wind.

The areal and stratigraphic distribution of the loess in North America thus seems to find a consistent explanation as a dual deposit made largely during the glacial ages by strengthened winds, from sources in arid regions, and in glaciated regions while outwash was being actively deposited.

Loess in Europe[59]

Areal distribution. The distribution of loess in Europe is shown on the generalized map (Plate 4). Basically the distribution is similar to that in North America, because loess is most abundant in the east, which has a plain and prairie terrain and a steppe climate, and is least abundant in the west, which has a dominantly hilly terrain and a maritime climate. As in North America it is apparent that loess is related to dry climates, to the border of the glaciated region, and to extensive outwash valley trains along the principal rivers. The vertical range of the loess approaches 2000 feet, so far up the flanks of the highlands has it lodged.

As in Iowa and Nebraska the loess in European Russia is continuous over wide areas and is relatively thick: sustained thicknesses of 30 to 50 feet continuous over thousands of square miles are found. It is thickest in the region west of the Dnepr and gradually thins out to the eastward. Farther west the loess breaks up into smaller, discontinuous areas, related principally to outwash masses and to highlands against which it has lodged. Westward too the thickness diminishes, commonly reaching only a very few feet but rising to 20–40 feet along major streams and approaching 100 feet along the east flank of the Rhine valley, its finest development in western Europe.

The outwash origin of loess in Europe was first recognized in 1889, and this origin is regarded as valid for the bulk of the European loess by most geologists. The close association of loess with outwash is clearly apparent in Germany. In France this association is seen along the Rhône and Garonne rivers, which carried outwash from glaciers in the Alps and Pyrenees respectively. In northern France the relationship to outwash is much less evident. Probably the "Channel River" was partly responsible:

[59] A convenient general reference is Grahmann 1932. See also Woldstedt 1929, pp. 111–124; Soergel 1919.

the relationship of loess to the present coasts of Brittany and Normandy suggests this. It has been held that, even where there is no outwash, spring floods would have been more sudden and abundant during the glacial ages than under present European climates,[60] and that this should have resulted, in major valleys, in the spreading out of much fine sediment for seizure by the strong cyclonic winds of the glacial-age spring seasons. This source may have supplied much of the loess in France.

In western Germany the relation of the loess to the Rhine, with its great train of Alpine outwash, is much like that of the loess in western Illinois, Kentucky, and Tennessee to the Mississippi. The loess is much thicker and more widespread east of the Rhine than west of it. The inference is drawn that westerly winds, the cyclonic planetary winds like those of today, were the chief agents of transport. In north-central Germany, on the other hand, the loess came chiefly from outwash in and around the *Urstromtäler* and was transported by northeasterly winds. Farther east loess distribution was influenced increasingly by northeasterly winds related to high atmospheric pressure over the ice sheet. The northeastern slopes of the Silesian highlands were mantled with loess from the great outwash areas of central Poland.[61] Throughout this broad region, however, loess is thicker on west-facing slopes than on slopes facing east. This suggests that loess was partly reworked by westerly winds traveling across Europe between the times of outbursts of cold air from over the Scandinavian Ice Sheet.

In south-central Europe the chief source of the loess was outwash in the valley of the Danube, derived from glaciers in the Alps. Silt from the upper Danube was blown back toward the Alps by northeast winds. From central Hungary, where Danube outwash was spread out wide, silt was picked up by southeast winds and deposited on the slopes of hills and mountains in Czechoslovakia, and was carried by northwest winds across Hungary into northern Yugoslavia. From the lower Danube silt was spread over eastern Rumania and the southern part of the Ukraine.[62]

A more important source of the widespread loess in southwestern Russia consisted of the great volumes of outwash deposited along the valleys of the Bug, Dnepr, Don, and Volga rivers. As is true of the Mississippi and the Missouri, the loess is thicker (in places more than 150 feet) in the vicinities of the Russian rivers than far from them.

[60] Grahmann 1932, p. 20.

[61] Conspicuously shown in Woldstedt 1935*b*, fig. 30.

[62] This is the view of Penck (1936*a*). Grahmann (1932) attributes this loess in part to local sources, on the ground that Danubian outwash was insufficiently bulky to account for all of it.

Scandanavia and Finland, like central and eastern Canada, are almost free of loess. As in North America, this is explained partly by a succession of great water bodies that followed the shrinking ice sheet in the Baltic region, partly by moist climate, and partly by bedrocks that do not yield abundant silt.

In the Mediterranean region there is very little loess, because of a conspicuous scarcity of outwash. The chief accumulations are in northern Italy (derived from the outwash in the Po valley) and in southern France (derived from outwash spread out by the Rhône). In the British Isles loess is little developed. Outwash is there, but it is not present in great volume, and the moist maritime climate may have maintained a carpet of vegetation over the valley trains even while they were actively accumulating, thus greatly reducing the areas of dry silt exposed to the wind.

In summary, the areal distribution and the variations in thickness of the European loess indicate that it was derived from widely and irregularly distributed sources, and that it reached its present distribution through the agency of winds blowing in various directions, whether they were the prevailing winds or not. In Europe the position of an area of loess deposition with respect to the position of a source of silt appears to have been a more important factor than the directions of the prevailing winds.

Stratigraphic position. Stratigraphically also the positions of the loess sheets in Europe resemble those in North America. Just as exposures of loess are found overlying each of the till sheets in central North America, so in Germany loess sheets have been identified on the Elster, Saale, and Warthe tills, and in the northern and western Alps region loess lies on outwash of both Mindel and Riss stages. The ancient Günz deposits are so thoroughly eroded and so poorly exposed that post-Günz loess, like post-Nebraskan loess in America, could formerly have been widely present, and yet be rarely apparent today.

The Würm drift in the Alps and the Weichsel drift in north-central Europe do not have loess covers. As in North America this absence of loess probably reflects the fact that not long after the Weichsel maximum great bodies of water succeeded each other in the Baltic region, forming traps for the sediment that would otherwise have been deposited on outwash plains. Meanwhile the *Urstromtäler* and other former sources, deprived of annual spring floods bearing heavy loads of sediment, became covered with vegetation and hence ceased to furnish silt to the winds.

In European Russia the loess, thicker and more widely distributed than farther west and thus reflecting the influence of the steppe climate,

is present on each of the three recognized drift sheets: the Likhvin, the Dnepr, and the latest drift sheet (including both Warthe and Valdai).

Loess in Siberia

In Siberia, according to Obruchev,[63] loess covers the steppe region north and east of the Caspian Sea in the Kazakh, Uzbek, and Turkmen republics, extending as far east as Lake Balkhash, and south over the foothills of the Pamir-Alai and T'ien Shan mountain groups. This vast area is adjacent to sources of silt in the exposed floors of former glacial lakes and in fine outwash sediments brought in by the Volga on the north and by streams draining the Pamir, Alai, and T'ien Shan mountains on the south.

Elsewhere in Siberia areas of loess occur in the Minusinsk basin south of Krasnoyarsk, in the Irkutsk region, in the region southeast of Lake Baykal, in the Vilyuy River-Lena River area, and along the lower Aldan River. Most if not all of these areas are along streams that carried outwash during the glacial ages. For this reason and others, "most Russian students believe that loess in European Russia and Siberia is a direct consequence of Pleistocene Glaciation."[64]

Loess in Other Parts of the World

Loess is not abundant in either Africa or Australia. It may be significant that, although arid regions cover wide areas on both continents, neither continent was more than slightly glaciated.

The eastern part of the South Island of New Zealand has extensive loess, reworked from abundant outwash carried eastward from the glaciers in the mountains along the west coast.

The plains region of Argentina and Uruguay, especially in the region between latitudes 30° and 40°, has thick and widespread loess, forming a broad belt northeast of the extensively glaciated country. Thicknesses are said to reach 30 to 100 feet over wide areas and in places to exceed 300 feet. But, because study of the loess has not been carried as far as in North America or Europe, eolian silt has not everywhere been clearly distinguished from waterlaid silt. Further, although more than one sheet of loess has been identified, the stratigraphic relations are not yet as well understood as they are elsewhere. There is difference of opinion as to the extent to which the loess represents reworked outwash; by some the loess is considered to be in large part the product of deflation of nonglacial alluvium from the semiarid country to the west, in the immediate rain shadow of the Andes.

[63] Obruchev 1945, p. 256.
[64] Obruchev 1945, p. 257.

Summary

The facts concerning the loess appear to add up to the conclusion that wind-blown silt has accumulated, near the maximum of each glacial age, at the periphery of each of the great ice sheets wherever the bedrocks, climate, and distribution of the drift were favorable. In general the favorable regions were beyond the southern and western sectors of the Laurentide Ice Sheet, the southern and eastern sectors of the Scandinavian Ice Sheet, and the northeastern sector of the ice in southern South America. The development of loess around the Siberian Ice Sheet is not yet well known.

A large amount of loess occurs in and to leeward of desert basins, particularly in Asia, but, as its source is mainly nonglacial alluvium, it is discussed only briefly.

It is doubtful that any consistent climatic significance can be attached to loess. The fact that loess is accumulating today under steppe climates, and the scarcity of loess in moist climates, together suggest that dry conditions are optimum for loess. But it seems probable that the continual deposition of fine alluvium such as characterized the major proglacial rivers, coupled with strong winds, would have resulted in the transport of great quantities of silt by the wind regardless of climate. The snail fauna of the North American loess suggests a climate not unlike that of the present day; the rodent fauna of the European loess suggests both arctic tundra and steppe conditions. As already indicated in at least one locality there is evidence of gradual change of climate during the accumulation of a single body of loess. In view of these facts, it does not seem possible to associate any particular climate with the times of great loess accumulation.

Chapter 11

GLACIAL STRATIGRAPHY

RELATION OF GLACIAL STRATIGRAPHY TO PLEISTOCENE STRATIGRAPHY IN GENERAL

The stratigraphy of the Pleistocene series involves the same range of deposits, the same kinds of problems, and the same techniques as the stratigraphy of older deposits, with the important addition that the direct and indirect effects of glaciation are more obvious than in any older group of strata, and that with terraces and similar features land forms have to be taken into consideration. Sediments of marine, brackish, lacustrine, fluvial, and eolian origin are widely present in addition to those of glacial origin. In some regions, indeed, Pleistocene volcanic rocks are abundant, and in others warped, folded, faulted, and greatly upheaved Pleistocene strata testify to the strength of Pleistocene mountain-making movements.

Although there are local exceptions, the Pleistocene nonglacial strata have generally been given little attention by geologists. In consequence we possess less information about them than about the glacial deposits. Further, glacial geologists have rather generally failed to consider the relations between the glacial record and the contemporaneous nonglacial record, and the converse has been almost equally true. As a result the study of the Pleistocene as a whole has suffered, because comparison of the glacial and nonglacial records has great potential value for strengthening the probability of correlations wherever climate plays a part. Accordingly, in the chapters that follow, this comparison is emphasized, though information is so scanty that hardly more than a beginning at correlation can yet be made.

As the principles of the stratigraphy of nonglacial deposits are well established, no special account of them seems necessary here. On the other hand glacial stratigraphy has certain peculiarities. These must be mentioned before we attempt to consider Pleistocene stratigraphy as a whole.

PECULIARITIES OF GLACIAL STRATIGRAPHY

The successful interpretation of glacial deposits, like the interpretation of strata of other kinds, rests basically on what William Smith in 1799 succinctly called *the order of superposition of the strata.* The deposits

underneath are older; those above are younger. Most glacial deposits, however, are thin, and in a mountainous region a valley glacier and its proglacial meltwater stream may mantle low terraces with deposits and yet fail to cover higher terraces that nevertheless had been covered with deposits of an earlier and much thicker glacier. If the low terraces were the product of an erosion interval separating the two glaciations, the older deposits might nowhere be in contact with the younger, and the younger would lie at a lower *altitude* than the older. Their relative ages would have to be inferred from an examination of other features, perhaps outside the immediate area. This general situation is common and frequently has to be reckoned with in glacial deposits, as in deposits of all other kinds controlled in part by the present topography.

Terrestrial deposits of whatever age are well known to be more difficult to interpret than marine deposits, mainly because of their greater lateral and vertical variability and their smaller content of fossils. Glacial deposits are primarily terrestrial, but they are more complex even than the general run of terrestrial sediments, for they are more variable and less well provided with fossils. The glacial strata, viewed broadly, include sediments laid down by ice, by streams, by the wind, by lakes, and by the sea. Also they are marked by erosional unconformities of unusual kinds. They exhibit curious stratigraphic relations caused by the rise and fall of sealevel and of lake levels, as well as by contemporaneous warping of the Earth's crust resulting from the loading and unloading of the great ice sheets. The fossils they contain, though few and far between, include plants and vertebrate and invertebrate animals, and both animals and plants include microscopic kinds. No wonder glacial sediments are varied and comprehensive. In volcanic areas such as Iceland, they even include interbedded igneous rocks.

In comparison with this complexity the tracing of a Paleozoic marine limestone seems almost simple. For such a task continuous or even very closely spaced exposures are not necessary because lateral variation is so gradual that continuity between exposures is reasonably certain. In contrast, the details of a single exposure of drift may be so complex, lateral variation so abrupt, and variations in thickness so great, that there can be no assurance of continuity without continuous exposure.

It is not even possible to correlate sections of till throughout a wide region on a basis of lithologic character, because the composition of a till changes with changing composition of the underlying bedrock. For this reason also successive tills in any one district tend to be much alike in composition.

Another complicating factor is the unconsolidated nature of most of the drift. Because of this, soil creep, slopewash, and slumping quickly

conceal an exposure, and vegetation soon covers it up. Hence the exposure can not be revisited at the frequent intervals that are possible with bedrock exposures, with any assurance that the relations it exhibits will still be visible.

Such specific difficulties aside, the interpretation of the glacial drifts and the nonglacial sediments related to them proceeds on normal stratigraphic principles.

DIFFERENTIATION OF THE DRIFT SHEETS[1]

The most important single result of glacial-stratigraphic study is the recognition of repeated glaciation. Only for a short time after the glacial theory took form could the Ice Age be thought to have consisted of a single glaciation. For in the 1850's evidence of two glaciations separated by ice-free conditions was recognized in Wales,[2] in Scotland,[3] and in the Alps.[4] Further investigations, detailed in a later part of this chapter, have led to the recognition of four distinct glaciations in North America and at least three in Eurasia. The factual basis of this standard subdivision is fourfold. It consists, in order of decreasing importance, of these elements: (1) degree of decomposition of the drift; (2) presence of nonglacial sediments between drift sheets; (3) degree of erosion of the drift; (4) recognition of the sources of stones in the drift. These four lines of evidence will now be considered.

Decomposition[5]

Perhaps the most satisfactory demonstration of repeated glaciations separated by long ice-free intervals lies in the gumbotils and related decomposition products that have developed in the clay-rich till sheets of the Mississippi basin. Each gumbotil clearly demonstrates a prolonged interval of weathering, and the fresh till overlying it records a subsequent glaciation.

These relations are shown in Fig. 47, a composite section in north-central Iowa where study of such features has been most intensive. The Nebraskan till, about 100 feet thick, rests on bedrock. The till consists of stones of many kinds in a very abundant matrix of clay and silt. The clay, derived from widespread shale bedrocks, predominates throughout the entire region. Because of it the till is comparatively impermeable to

[1] See Salisbury 1893*a;* Bain 1898, p. 25; Blackwelder 1931; Thwaites 1928.
[2] A. C. Ramsay 1852.
[3] Chambers 1853.
[4] Morlot 1856.
[5] See a good summary in Thwaites 1941, pp. 62–65.

circulating subsurface water, so that subsurface drainage is poor and percolation is slow.

Where the surface of the ground is broad and flat the upper part of each of the two older tills consists of *gumbotil*,[6] a gray, leached, deoxidized clay, consisting chiefly of the mineral beidellite. It is very sticky

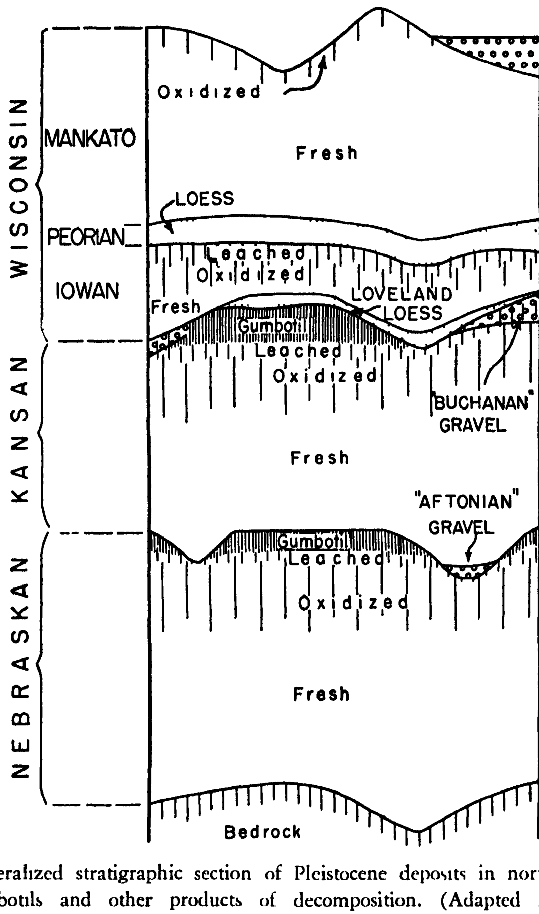


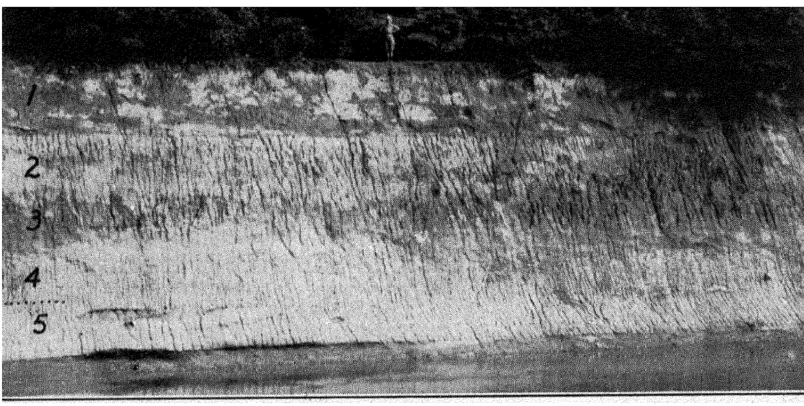
FIG. 47. Generalized stratigraphic section of Pleistocene deposits in north-central Iowa, showing gumbotils and other products of decomposition. (Adapted from Kay and Apfel 1929.)

when wet and extremely firm when dry. The clay contains stones, but most of them are of silicious types highly resistant to decomposition. Counts made in Iowa show that in the Kansan gumbotil 87 per cent of the stones are silicious, whereas in the zone immediately beneath the

[6] Kay and Pearce 1920. The word is an adaptation of the popular word *gumbo* — a highly descriptive name, as anyone will understand if he has tried to drive a car on a non-surfaced road in gumbotil country on a rainy spring day.

gumbotil only 42 per cent of the stones are silicious. The inference is plain that the nonresistant stones in the top of the till have been destroyed by decomposition. Actually some exposures of the base of a gumbotil show a transition zone in which the faint outlines of almost completely decomposed stones are still visible.

Beneath the gumbotil is a zone in which the till has been oxidized and also leached of its content of carbonates. This zone grades downward into a thick zone in which the till is oxidized but not leached. Finally, beneath this, is fresh unaltered till.



Paul MacClintock, Illinois Geol. Survey

FIG. 48. Exposure of gumbotil at the top of Illinoian till, Embarrass River, Illinois. 1, loess (Peorian); 2, gumbotil; 3, oxidized and leached till, with concentration of iron oxides; 4, oxidized and leached till; 5, oxidized till.

The whole vertical sequence of zones, termed a *soil profile* (Fig. 48) because it was first recognized by soil scientists attempting to determine the origin of soils through chemical alteration of the bedrocks, conforms to a standard. This standard is represented in various ways, of which the one offered by Leighton and MacClintock[7] is best adapted to use in glacial geology. It consists of five zones, numbered from the top downward:

ZONE

1. Surficial soil.
2. Drift, chemically much decomposed.
3. Drift, oxidized and leached of carbonates but otherwise little altered.
4. Drift, oxidized but containing primary carbonates.
5. Drift, unaltered.

[7] Leighton and MacClintock 1930.

Beginning at the bottom and going upward, we can trace the succession of changes that progressively affected the drift from the top down. (1) *Oxidation* is the pioneer reaction, taking place first and extending deepest. Ferrous iron compounds are altered to oxides whose red, yellow, and brown tones impart a distinctive appearance to the section. (2) *Leaching* of the primary carbonates in the drift, by percolating water charged with carbon dioxide, follows close behind the oxidation process. In time, leaching dissolves not only the finely divided carbonates (mostly pulverized limestone and dolomite) in the till matrix but entire stones and boulders of carbonate rocks as well. (3) *Complete decomposition* of the drift follows slowly as percolating water attacks and alters even the stubborn, little-soluble silicates to "clay minerals," consisting principally of hydrous aluminum silicates. These are deposited little by little at ever deeper levels, as decay of the primary minerals above causes the surface of the ground to subside slightly. The "clay minerals" form the sticky gumbo and, together with whatever resistant quartz is present, constitute the residual end product of the alteration process. As stated in Chapter 18, probably at least 100,000 years are required for the development of a thick layer of gumbotil on a mass of tight, clay-rich till.

In areas of steeper slopes and in silt-rich till, more permeable to water than clay-rich till, water circulation is more rapid and the alteration process speeds up. The resulting end product, *silttil,* has the texture of gritty silt rather than sticky gumbo, chiefly because the silt in the original till consists mostly of quartz.

In very sandy till and in sand and gravel, permeability is so great that water moves through them quickly. In such material gumbotil can not develop. Instead, there results a deep zone of oxidation colored brownish, reddish, or yellowish. Little leaching is evident because such material is usually very silicious with little primary carbonate.

As the stones and boulders in the deposit are altered progressively from their surfaces inward, they acquire *weathering rinds* that are oxidized to a brownish color and are otherwise decomposed.

In Fig. 47 it is apparent that the Iowan till is not capped with gumbotil. The alteration process on that till has not progressed beyond oxidation and leaching. The overlying Mankato till has not even been leached appreciably. A thin oxidized zone, forming a *skeleton soil profile,* represents its maximum degree of alteration.

In summary, repeated glaciation is recorded by fresh till overlying decomposed till, and the degree of decomposition gives some indication of the relative lengths of time that elapsed between successive glaciations. The time intervals can not be determined accurately, because the rate

of decomposition depends on factors that are variable. Chief among them are climate, topography, permeability of the drift, and chemical composition of the unaltered drift.[8] A till consisting chiefly of limestone grains of silt size, in a hilly region and under a moist climate, would decompose more rapidly than a till consisting chiefly of clay in a plains region with a semiarid climate.

In permeable types of till, decomposition is likely to be spread through so deep a zone that the elements of the soil profile are indistinct. In these situations fairly successful attempts have been made to separate drifts of two or more different ages on the basis of the extent to which decomposition has penetrated into the contained granitic stones.[9]

Some tills contain a high proportion of decomposed material that had been picked up by the flowing glacier and mixed with inconspicuous amounts of fresh debris before deposition. Such tills, whether or not overlain by younger deposits, give a false impression of age because of the decomposed material they contain. Close inspection shows that the decomposed elements extend through the deposit instead of being related to a soil profile. Microscopic examination establishes this fact clearly, by revealing fresh grains of mineral types readily susceptible to decomposition, side by side with thoroughly decomposed grains of the same types.[10]

Sections exposing two tills without intervening decomposition are numerous, but they do not prove multiple glaciation. They may represent fluctuation of the position of the glacier terminus through very short periods. In some sections the contact between the tills is marked by a concentration of stones or boulders — a *boulder pavement*. The pavement is generally interpreted as a lag concentrate and is thought to indicate erosion of the upper part of the older till before deposition of the younger. Under these circumstances a zone of decomposition at the top of the older till might have been removed without trace, and it is therefore not possible to estimate how much time elapsed between the two episodes of till deposition.

Nonglacial Sediments

A second kind of evidence of repeated glaciation consists of nonglacial sediments lying between sheets of till and, less satisfactorily, between deposits of stratified drift. In this category the most reliable occurrences consist of two tills separated by sediments that contain fossil plants or

[8] See Alden 1909 for an elaboration of some variable factors.

[9] Blackwelder 1931; MacClintock 1940; Page 1939, p. 790.

[10] See, for example, Flint 1937, p. 215; Krynine 1937. A case involving other "false" indications of age is discussed in Hole 1943.

animals characteristic of a nonglacial climate. Commonly the sediments consist of peat. Because they consist largely of the remains of trees, these peat layers were formerly known, both in Britain and in America, as *forest beds*, but this term has been falling into disuse.

In both North America and Europe fossil spruce and fir are very common in the inter-till sediments. These trees, characteristic of the subarctic forest, could have lived in fairly close proximity to the great ice sheets and hence do not prove extensive deglaciation. In fact it has been argued[11] that such vegetation might have flourished on ablation moraine still underlain by glacier ice, though this seems unlikely save in regions of pronounced maritime climate.

On the other hand deposits such as those (described elsewhere) at Toronto, Hötting, and Eem, all of which contain fossil species characteristic of climates warmer than the present climates at those places, probably indicate very widespread deglaciation between two glaciations. The inter-till sediments are of several kinds — bog peats, alluvium, lake sediments and sea-floor deposits — but the nature of the sediments is far less important than the ecology of the fossils they contain.

Among sea-floor sediments that record multiple glaciation is the Gardiners "clay" (= Cape May formation) on Long Island, a marine silt lying between tills and containing fossil invertebrates that indicate a mild climate.[12]

Other nonglacial sediments include loess, which, though extensive, proves little as to the extent of deglaciation because it is confined to the peripheral areas of the drift sheets. Again, in some places stratified drift not containing fossils lies between two tills. Unless well-developed soil profiles are involved, such occurrences mean little because a section consisting of two tills separated by stratified drift could be developed during a single glaciation, either through slight fluctuation in the position of the margin of the glacier or through the development of segregated "melts"[13] within a glacier that is not fluctuating but is steadily wasting away.

Extent of Erosion

Multiple glaciation is indicated by extent of erosion in two different ways. The more definite consists of evidence that deep and widespread erosion of bedrock occurred between the times of deposition of two tills.[14] Examples are widely present in the Rocky Mountains in

[11] Tarr 1909*a*, pp. 104–105.
[12] Discussed in Chapter 14.
[13] R. G. Carruthers 1939.
[14] Atwood and Atwood 1938.

Montana[15] and in the San Juan Mountains in Colorado.[16] In both districts the intervening erosion of bedrock amounts in depth to many hundreds of feet. Even were it not confirmed by decomposition of the older till, a very long time interval is demanded by this deep erosion.

Less definite is evidence consisting of difference in degree of erosion of two till sheets in the same region. Such difference is evident in central Illinois. There the Illinoian drift, where not overlain by Wisconsin drift, has subdued topography with few closed depressions and with slopes much modified by mass-wasting. The overlapping Wisconsin drift, with similar composition, has more pronounced constructional topography with steeper slopes and many closed depressions. The difference between the two is clearly evident and involves a long interval of erosion, though it provides no basis for a satisfactory estimate of the actual amount of time involved.

Provenance

Under favorable conditions, and where the bedrock geology of the region has been thoroughly studied, it is sometimes possible to infer, from lithologic differences between two tills, a long time interval between them. In the San Juan Mountains in Colorado,

> During the time which intervened between the deposition of the Cerro till in the Uncompahgre Valley, with its paucity of quartzite boulders, and the Durango glaciation in that same valley, the Uncompahgre Canyon south of Ouray must have been carved deeply enough into the quartzite of the Uncompahgre formation to expose that formation so generously that the Durango ice could pluck an abundance of boulders from it. Similarly, in the Blanco Valley on the south side of the range the stripping of the cover of volcanic rocks from the surface of the monzonite-porphyry laccolith must have been largely accomplished in the post-Cerro and pre-Durango interglacial interval, for boulders of the monzonite are absent from the older till but numerous in the later drift.[17]

Summary

Review of the evidence of multiple glaciation reveals that the best indications are (1) fresh till overlying deeply decomposed till, and (2) two tills separated by nonglacial deposits containing fossil plants or animals that record climates at least as mild as the existing climates. Other kinds of evidence exist but for various reasons are less satisfactory.

[15] Alden 1932, p. 32.
[16] Atwood and Mather 1932, p. 28.
[17] Atwood and Mather 1932, p. 129.

TERMINOLOGY OF POST-PLIOCENE STRATIGRAPHY

Introduction

The terminology of post-Pliocene time, its equivalents and its subdivisions, has been in process of evolution since 1822. During the first half of this time the terminology evolved in conformity with growing knowledge. During the second half, although knowledge has grown even more rapidly and has led to the adoption of a recognized sequence of glacial and interglacial ages, the fundamental classification of post-Pliocene time has remained unchanged. Accordingly, because we are now able to look realistically at a classification that no longer meets the facts, the time has come when the classification should be modified. Most stratigraphers and glacial geologists realize the weakness of the present classification, but no firm proposal for modification has been made. To suggest a reasonable modification is the primary purpose of the discussion that follows. Without a firm and logical basis of stratigraphic terminology it would be very difficult to write the remainder of this book. Specifically, the present fundamental classification commonly current in the English-speaking world is as follows:

Cenozoic (Kainozoic) era	Quaternary period	Recent epoch Pleistocene (= "Glacial") epoch
	Tertiary period	Pliocene epoch and earlier epochs

Historical Basis of Present Usage

The scheme now current is essentially of mid-nineteenth century date. Its evolution began with the introduction of the term *Quaternary* by Desnoyers[18] as an addition to the standard designations *Primary, Secondary,* and *Tertiary,* all of which had been in use since 1760. Desnoyers applied the term (in a curiously tentative way) to the strata that overlie the Tertiary rocks in the Paris basin. In the following year Reboul[19] extended the term to include the strata characterized by living species of animals and plants, as distinguished from the Tertiary, whose fossil remains, he thought, "almost all belong to extinct species."[20] Reboul's definition, we may note, was the first one to have a faunal basis. He subdivided the Quaternary as follows:

Quaternary period	Historic epoch Anti[*sic*]-historic epoch

[18] Desnoyers 1829, p. 193.
[19] Reboul 1833.
[20] That this belief was not soundly based was being shown, across the English Channel, in that same year by Lyell (1833, pp. 52–58), who was just then introducing the terms *Pliocene* and *Recent.*

In a paper read in 1854 but not published until two years later Morlot[21] subdivided the Quaternary ("Quartaire") into

$$\text{Quaternary period} \begin{cases} \text{Modern epoch} \\ \text{Second glacial epoch} \\ \text{Diluvian epoch} \\ \text{First glacial epoch} \end{cases}$$

and restricted it to the post-Pliocene. This classification was glacial rather than faunal. Incidentally its recognition of distinct glacial and interglacial times shares with that of A. C. Ramsay in Great Britain the distinction of being the first on record.

While Reboul in Paris was redefining the Quaternary, Lyell[22] in London introduced the term *Recent epoch,* defined as the time "which has elapsed since the earth has been tennanted by man," and recognized as younger than the Tertiary. As thus defined it included both the Pleistocene and the Recent of the classification commonly current today.

Six years later Lyell[23] introduced the term *Pleistocene,* applying it to strata whose content of fossil mollusks included more than 70 per cent of living species. Defined in that way, the Pleistocene included a part of the Tertiary as now defined. But shortly afterward Forbes[24] redefined the Pleistocene epoch as equivalent to the "Glacial epoch"[25] — "the time distinguished by the presence of severe climatal conditions through a great part of the northern hemisphere." Although Forbes thought of the "Glacial epoch" as a time of vast iceberg-laden seas rather than of great ice sheets, nevertheless he is the originator of the definition of the Pleistocene in terms of climate. Lyell nevertheless continued to regard the Pleistocene as synonymous with post-Pliocene regardless of whether it included only the glaciation or more than the glaciation.[26]

Forbes' removal of the Pleistocene from the Recent, agreed to by Lyell in 1873, automatically reduced the Recent (also sometimes called Holocene) to the equivalent of the postglacial.

This brings us back to the current meanings of these terms. As defined by the U. S. Geological Survey[27] the Quaternary period includes the Pleistocene and Recent, is characterized by animals and plants of modern types, and is commonly called the "Age of Man." The Pleistocene ("Glacial") epoch, also popularly called the "Ice Age" or "Great Ice

[21] Morlot 1856.
[22] Lyell 1833, p. 52.
[23] Lyell 1839, p. 621.
[24] Forbes 1846, p. 402.
[25] A *Glacial period* (*Eiszeit*) had been recognized in 1837 by Karl Schimper (1837) — also, significantly, in Switzerland where glacial evidence is abundant.
[26] Lyell 1873, p. 3.
[27] Wilmarth 1925, p. 45.

Age," includes the extensive glacial deposits[28] of the northern hemisphere and contemporaneous rocks. The Recent epoch includes the deposits[28] made since the disappearance of the former great ice sheets.

In France, a country without extensive glacial deposits, the tendency has been to continue to define and subdivide the Quaternary on a basis of fossils. Compare Haug,[29] who regarded the Pleistocene as beginning with the appearance and intercontinental spreading of three new mammals: *Bos, Elephas* (in the broad sense), and *Equus*. This view has been recently urged by Hopwood and others[30] and is further discussed elsewhere in this chapter.

In German usage, the classification is the same as that current in the English-speaking world, though the terminology differs. For Pleistocene the Germans use *Diluvium,* and for Recent, *Alluvium*. Curiously enough this terminology had its origin not in Germany but in England, apparently with Mantell,[31] who classified the superficial deposits into Diluvium (described as sediments laid down by agencies no longer operative, such as the biblical flood) and Alluvium (sediments laid down by agencies still in force, such as existing streams). In German usage Diluvium (=*Eiszeit*=*Glazialzeit*) and Alluvium (=*Postglazialzeit*) are grouped together to form the Quatär (=Quartär=Quatärzeit) in exact analogy with English-speaking usage.

A modification of the generally accepted usage was proposed by Hershey.[32] He recognized a long interval of erosion following the Pliocene and preceding the first invasion by the great ice sheets. This he called the "Ozarkian epoch." A similar scheme was urged by LeConte,[33] who placed the "Ozarkian epoch" after the Pliocene, regarded it as the last epoch of the Cenozoic, and considered the subsequent "Glacial epoch" to mark the beginning of a new era, the "Psychozoic era." These schemes have not received general support. Hershey thought the "Ozarkian" interval amounted to at least half of post-"Lafayette" (Pliocene, Brandywine) time. Using as a basis fossil mammals, Osborn[34] restricted Glacial time to the midpart of the Pleistocene epoch, which he regarded as including also a Preglacial and a Postglacial, and as being followed by a Recent or Holocene epoch. Later Eaton[35] revived the view

[28] Evidently the time of the deposits is meant rather than the deposits themselves. The correct usage is outlined at the end of the present discussion.

[29] Haug 1911, p. 1776.

[30] Cf. Hopwood 1935, p. 47; Paterson 1940.

[31] Mantell 1822, p. 274.

[32] Hershey 1896.

[33] LeConte 1899.

[34] Osborn 1915, p. 218.

[35] Eaton 1928; 1941.

that as measured by nonglacial sediments and fossils the Pleistocene epoch endured long and that glaciation, at least on the basis of evidence in California, did not begin until well on in this time unit. He stressed the possibility that the four glaciations of which we have direct evidence are only the later members of a series of which the earlier are of lesser extent and therefore not easily distinguishable. This is a possibility that should not be overlooked because it points the way toward renewed attempts to determine the age relations between glacial and nonglacial Pleistocene in various parts of the world.

In summary, our generally used classification consists of a Quaternary period embracing a Pleistocene epoch and a Recent epoch. The basis of this classification needs to be re-examined. First we will examine the nature of the Pliocene-Pleistocene break in order to determine whether it is so distinct and conspicuous that it can properly be considered as separating two systems of rocks from each other.

Importance of the Break between Pliocene and Pleistocene

Throughout most of the geologic column the breaks between successive systems of strata (representing periods of geologic time) are marked by strong deformation of the rocks in important orogenic zones, profound erosion, and major changes in faunas and floras. It is because they possess these major features that such breaks are regarded as having "period value"; that is, they are thought of as being second only in rank to the breaks that separate geologic eras.

When the Pliocene-Pleistocene boundary is examined in this light it does not measure up to the stature of the boundaries between other periods. Wherever Pliocene and Pleistocene strata have been investigated in detail, there has been a tendency to regard the relationship between them as transitional. Neither physically nor in the organic record does the boundary seem to be an important one.

In East Anglia the boundary is so inconspicuous that the upper part of the crag sequence (Table 11) has been variously regarded as upper Pliocene and lower Pleistocene. Opinion is apparently now tending toward the latter view.[36] Similar inconspicuousness in the Paris basin was clearly recognized by Desnoyers when he coined the term *Quaternary*.[37]

In the Mediterranean region, similarly, the Calabrian marine sequence has been thought to represent both the youngest Pliocene and the oldest Pleistocene, although the present tendency seems to be to consider it wholly Pleistocene. The terrestrial equivalent of the Calabrian, the Villa-

[36] Zeuner 1937.
[37] Desnoyers 1829, p. 193.

franchian, formerly widely regarded as Pliocene, has been placed in the Pleistocene by several authorities.[38]

In eastern United States neither late Pliocene nor early Pleistocene marine deposits have yet been recognized; hence the Pliocene-Pleistocene contact in that region is not of much aid in determining the character of the boundary.

In the terrestrial strata in the Great Plains region of west-central North America the boundary is not conspicuous, and in places at least the Pliocene-Pleistocene sequence is transitional. Opinions differ as to what should be considered the base of the Pleistocene, and no agreement has been reached.[39] Further, the freshwater and terrestrial mollusks in central North America show no conspicuous change from Pliocene to Pleistocene.[40]

In the northwestern interior of North America the terrestrial deposits constituting the Idaho formation have fossil plants with a Pliocene aspect and fossil animals with a Pliocene and Pleistocene aspect.[41]

In southern California, where there is a thick Pliocene-Pleistocene sequence, the conspicuous physical break, marked by a profound unconformity, occurs far above the base of the Pleistocene according to every classification thus far proposed. If fossil invertebrates and vertebrates alone are considered, opinions differ so widely as to what should be considered the base of the Pleistocene that agreement has been thus far impossible.[42]

In the Himalayan region, likewise, the thick Siwalik terrestrial sequence records both Pliocene and Pleistocene deposition, and opinions have differed as to where to put the boundary between them. De Terra and Paterson[43] place it at a disconformity between Middle and Upper Siwalik, which they regard as probably, though not certainly, marking a long hiatus.

In South Africa the boundary between Pliocene and Pleistocene is regarded as vague.[44] In Tasmania there has been stated to be "no justification for separating Pliocene, Pleistocene, and Recent."[45] Glaciation has been regarded repeatedly as extending back into the Pliocene.[45a]

The foregoing examples represent most of the regions in which the

[38] Colbert 1942, p. 1431.

[39] Frye 1945, pp. 76, 90; Schultz 1938, pp. 95–98.

[40] H. G. Richards, *unpublished.*

[41] Kirkham 1931.

[42] Weaver and others 1944, pl. 1. See also Colbert 1942, p. 1505.

[43] De Terra and Paterson 1939, pp. 253–256. However, De Terra (1940, p. 121) stated that an angular unconformity marks the base of the Pleistocene throughout southern Asia.

[44] Du Toit 1939, p. 510.

[45] A. N. Lewis 1934, p. 70.

[45a] Cf. Gerasimov and Markov 1939; Gage 1945.

Pliocene-Pleistocene sequence has received detailed study. From the evidence they offer we may conclude that the boundary in question does not have "period value" and can not justifiably be considered as separating units of higher rank than epochs. From this conclusion it follows that the ancient concept of a "Tertiary period" and a "Quaternary period," though fully adequate for the time when it was first used, does not now rest on firm ground.

Because of the lack of a conspicuous physical break between Pliocene and Pleistocene in the Paris basin, Lyell separated the two units on the basis of their content of fossil mollusks. Strata with a molluscan fauna of which more than 70 per cent of the species are still living he called Pleistocene. Owing to refinements of study this percentage has been changed to about 90, but even with this revision the Lyell percentage system is unsatisfactory because it has been found to vary from region to region.[46] Faunally as well as physically therefore the Pliocene-Pleistocene contact is uncertain.

In summary it seems evident that a widespread physical break between Pliocene and Pleistocene, comparable with the breaks between systems in the geologic column, does not exist. The Pliocene-Pleistocene boundary is transitional in some sections and is at best inconspicuous. Faunally, the "percentage of molluscan extinction" used as a basis by Lyell has not proved to be widely useful. Mountain uplift continued from the Pliocene into the Pleistocene. The changes in world geography were not great enough to bring about a drastic change in the forms of animals and plants. Yet from every practical point of view the Pleistocene unquestionably deserves recognition as a separate unit. A sound basis for distinguishing it is needed.

Climate as the Basis of Distinction between Pliocene and Pleistocene

In the absence of the means of differentiation commonly used in older strata, the best basis of separating the Pleistocene from the Pliocene is unquestionably climate. The climates that resulted in extensive glaciation constitute the one outstanding and basic respect in which Pleistocene time differed from the time preceding it.[46a]

[46] Cf. Schuchert and Dunbar 1941, p. 384; H. R. Gale 1931, p. 20; Dickerson 1921.

[46a] After the present work had been completed, the author encountered an identical opinion, clearly stated by a paleontologist. The reader is referred to Schultz (1938*a*, p. 96) for a cogent argument which, among other things, furnishes suggestive grounds for extending the Pleistocene down to the base of what is now considered upper Pliocene in western North America.

> In every respect the Pleistocene epoch is allied with the Pliocene, and if it were not for the extensive glaciation that characterized the Pleistocene, the two would probably never have been differentiated.[47]

According to the present weight of evidence the great climatic fluctuations were simultaneous throughout the world, and their direct and indirect effects were profound. The major direct result was the burial (at the glacial maximum) of 32 per cent of the land area of the world beneath glacier ice, and the imprinting on most of this area of a conspicuous identifiable record. Another result, extending for hundreds of miles into the nonglaciated land areas, was outwash sedimentation, making possible the correlation of features in additional wide nonglaciated regions. A third result was the conspicuous change wrought by moist (pluvial) climates in the now desert and semidesert regions of middle and low latitudes, creating lakes and streams and altogether affecting another 10 per cent of the world's land area. A fourth result was the notable fluctuation of sealevel that took place in sympathy with the fluctuating glaciers. It affected all the coasts of the world and temporarily added (at the glacial maximum) 2 or 3 per cent to the land area of the world by exposing extensive areas of continental shelf. Finally, a fifth result consisted of widespread migrations of animals and plants, leaving at any one place a record of notably different assemblages at different times, not only on the nonglaciated lands but in the floor deposits of shallow seas as well.[48]

The existence and wide extent of all these related features makes a strong case for the use of climate as a basis for the definition (and also for the subdivision) of the Pleistocene, which should be thought of as beginning with the first evidence, direct or indirect, of the existence of a notably cool climate. On the other hand marine and terrestrial sediments in very low latitudes where the influence of climatic change was comparatively slight may demand some additional and more local basis for correlation, although it seems probable that as knowledge increases even these can be related in one way or another to contemporaneous events in higher latitudes.

It is true that the beginning of glaciation was not everywhere contemporaneous. Glaciers formed in high mountains long before they invaded lowlands in middle latitudes. But this fact does not constitute a

[47] Schuchert and Dunbar 1941, p. 411.

[48] That climatic inference should be looked for in nonglacial sediments as the readiest means of correlation with the glacial record has been clearly stated by H. R. Gale (1931, p. 21). Climate as an important element in the definition of the Pleistocene has been stressed by Movius (1944, p. 8), Zhirmunskiy (1931), and Schultz (1938*a*).

valid objection to the use of climatic fluctuation as a basis of correlation. Similar lack of time equivalence is a characteristic of the majority of stratigraphic basal horizons. The encroachment and regression of any Paleozoic sea constituted a gradual and progressive event. The time unit represented by its deposits began much earlier in some places than in others.[49] Yet the unconformity at the base of the deposits is accepted as marking the beginning of the new stratigraphic unit. The unconformity at the base of the earliest glacial deposits should be treated in the same way.

A different basis of distinction that may have much value in some regions and that may (though not necessarily) be independent of climatic fluctuation must be considered also. It was stated earlier in this chapter that Haug had suggested that the Pleistocene might be regarded as beginning with the appearance of *Bos, Elephas* (in the broad sense), and *Equus*. This suggestion has been approved by several paleontologists and has been extended to include the modern camel types and man.[50] If, as is quite possible, these forms should prove, through their relationship to the climatic evidence, to be peculiar to the Pleistocene and therefore true Pleistocene guide fossils, the fact would have great practical usefulness. The wide provisional adoption of this group of fossil forms as Pleistocene constitutes a logical argument in favor of incorporating the concept into any working definition of the Pleistocene.

The Terms Tertiary and Quaternary

It has been shown that no great and widespread unconformity separates Pleistocene from Pliocene strata, and that differences between the faunas of these two epochs, although evident, are far from large. The Pleistocene epoch differs from its predecessor chiefly in the fact of conspicuous climatic fluctuation.

If we compare this difference with the differences involved in the distinctions between other pairs of members of the stratigraphic column we are obliged to conclude that it can not logically have period value, but only epoch value. From this conclusion it follows that the distinction between a *Tertiary period* and a *Quaternary period* is invalid, because there is no geologic basis for drawing between them a line of period value.[51] When we add to this the fact that the terms *Tertiary* and *Quaternary* are parts of an abandoned general classification dating from 1760

[49] Incidentally the encroachment of a Paleozoic sea probably took much longer than the advance of a Pleistocene ice sheet over a comparable area.

[50] E. H. Colbert, *unpublished;* see also Colbert 1942.

[51] Schuchert and Dunbar (1941, pp. 383–384, 425) emphasized the undesirability of these terms.

with an addition in 1829, and that the designation *period* was used in only a vague general sense when this classification came into being, the conclusion is plain that (1) the terms Tertiary and Quaternary ought to be dropped from our nomenclature just as Primary and Secondary were dropped long ago, and (2) no periods ought to be recognized within the limits of the short Cenozoic era, although a Cenozoic period might properly be recognized as coextensive with the Cenozoic era up to and including the present time.

Pleistocene-Recent Boundary and the Term Postglacial

When we turn to the Pleistocene-Recent boundary we find again that opinions differ widely as to what should be taken as marking the end of the Pleistocene and the beginning of the Recent. The only character common to all the proposals is that they are arbitrary.

The earliest discrimination between the two, that of Forbes,[52] is a simple discrimination between glacial and postglacial. Unless applied to a small area, this basis of distinction is extremely vague, because the last deglaciation was progressive. It uncovered the peripheral areas tens of thousands of years before the central areas were freed from ice, and it left records of well-marked pauses in the process. Yet the distinction between Pleistocene and Recent is still widely used.

The present time differs from the time when the great ice sheets were last at their maxima in that many mammals then present have become extinct. The difference is noteworthy, but as extinction was progressive (just as deglaciation was progressive), it would be extremely difficult to apply in the recognition of a sequence of deposits. Although the extinction of these mammals probably constitutes a better basis than any other for recognizing a break between Pleistocene and "Recent," the disadvantages of recognizing such a break seem far to outweigh the advantages.

There is said to be no difference, except locally in range, between the marine and fresh-water invertebrate faunas of the "Late Pleistocene" and "Recent."[53] Only two or three of the "Late Pleistocene" species are extinct today.

As early as 1872 Reade,[54] recognizing the vagueness of the existing usage, suggested that the Recent be considered the time during which the last major rise in sealevel has occurred. This suggestion has been recently revived and indorsed by R. J. Russell.[55] As a basis of classification, rise

[52] Forbes 1846.
[53] H. G. Richards, *unpublished.*
[54] Reade 1872, p. 111.
[55] R. J. Russell 1940, p. 1201.

of sealevel has the advantage of being contemporaneous throughout the world, but it is extremely difficult to apply stratigraphically. It can not be recognized in the continental interiors, and along coasts the evidence, even where uncomplicated by crustal warping, is obscure and unsatisfactory. Therefore, though possessing great local value, rise of sealevel is hardly likely to become a generally used basis for marking the beginning of the Recent.

De Geer adopted the time when the shrinking Scandinavian Ice Sheet separated into two parts at Ragunda in central Sweden as marking the end of the "Glacial age" and the beginning of the "Postglacial." He did not use the terms *Pleistocene* and *Recent.* This event is a convenient arbitrary point in the Swedish chronology, but it is not applicable elsewhere.[56]

At the close of Lake Algonquin time, the opening of the North Bay lake outlet marked the end of the ice-dammed lake sequence and the beginning of the ice-free lake sequence in central North America. "This event," wrote Leverett and Taylor,[57] "may be taken as the place of division between the Pleistocene and Recent epochs of geologic time." This is a logical and clearly marked point in time, but it is wholly arbitrary and is not recognizable outside the Great Lakes region.

Morgan[58] attempted a twofold distinction. He adopted the time when the receding margin of the last Scandinavian Ice Sheet cleared the north German coast, and when the North American Ice Sheet disappeared from the lowlands of the middle latitudes, as arbitrarily marking the beginning of the Recent. Like the other arbitrary proposals, this scheme has only local application, and it is not even certain that the two named events were contemporaneous. The same author also defined the Recent paleontologically as the time whose mammals and other large animals, with few exceptions, are still living. This second definition not only is vague but moreover can not be correlated with the first one.

Antevs,[59] apparently following Enquist,[60] took the end of the Pleistocene as "the time when the temperature in the southern and greater part of the once glaciated area had risen to equal the present temperature," but he did not explain how the temperature rise could be recognized. His classification departs from common usage in that he considers the Pleistocene to be followed by the post-Glacial, which in

[56] De Geer 1912, pl. 1. Madsen and others (1928, p. 121) took the opposite view, that to set up a sharp boundary between glacial and postglacial, even in so small an area as Denmark, is impossible.

[57] Leverett and Taylor 1915, p. 331.

[58] Morgan 1926.

[59] Antevs 1931, p. 1.

[60] Enquist 1929.

turn is subdivided into Early, Middle, and Late. The Late Post-Glacial is further subdivided into Recent and Modern. This complex arrangement is cumbersome, and, in the absence of readily recognizable evidence of the temperature changes referred to, it does not seem workable.

The arbitrariness of all the proposals for separating postglacial (= Recent of many authors) from glacial (= Pleistocene of many authors) arises from the fact that the term postglacial is properly referable to only one district at a time. Events that took place while the district was last covered with glacier ice are referable to glacial time *for that district;* subsequent events are there referable to postglacial time. Since removal of the glacier cover was progressive, it follows that postglacial time began for some districts much earlier than for others, and that for most of Antarctica and Greenland it has not begun yet.

The term *postglacial* can properly be applied to events in regions that never were glaciated but in which the indirect effects of glaciation are clearly marked, as for example the valleys of large streams draining glaciated regions. The cessation of outwash deposition and the beginning of trenching of the outwash by nonglacial runoff is an event that can be definitely recognized in many places and that can be taken to mark the transition from glacial to postglacial time in a given valley system. Where there is no direct connection, such as is afforded by outwash, between a glaciated and a nonglaciated district, the term postglacial is hardly applicable to any feature in the nonglaciated district. For the desert basins of western United States, where Pleistocene features record alternating episodes of moisture and dryness, terms such as Pluvial and Postpluvial[61] are sometimes used.

There is thus a real objection to assigning to the postglacial the value of an epoch, in that it can not be satisfactorily recognized in the field as a time unit. If the evidence were obscure, like the evidence of the onset of Pleistocene glaciation, we would be justified in generalizing our interpretation of it, but it is so detailed and so abundant that we can not justifiably close our eyes to the gradual transition it records.

Another objection to assigning to the postglacial the value of an epoch is that in North America, at any rate, there were at least four glacial ages separated by three interglacial ages, in one of which the climate was warmer than the present climate. Further, whereas more than 30 per cent of the world's land area was covered by ice during the maximum glaciation and about 27 per cent was covered during the last glacial age, about 10 per cent still remains covered today. This present condition is hardly "postglacial," nor is it, we believe, characteristic of preglacial

[61] Cf. Antevs 1936.

time. The inference is clear that the term postglacial is of only local value and of only subordinate stratigraphic importance and should be used only in an informal way. As Daly[62] put it:

> Geology has been developed by men living in the middle latitudes of the Northern Hemisphere. To those leaders of thought the Glacial period is of the past. For a geologically-minded penguin of Antarctica the Glacial period is here and now. The truly objective philosopher must agree with the penguin: the earth as a whole is something like halfway through the Glacial period.

Definition and Subdivision of the Pleistocene

Since we propose, as Forbes did in 1846, to define the Pleistocene epoch and to set it off from the Pliocene principally on a basis of climate, the only logical course is to consider the Pleistocene as still in progress. Accordingly, for the purposes of this book, the Pleistocene epoch is provisionally defined as that part of late-Cenozoic time which is characterized by repeated climatic cooling, involving repeated conspicuous glaciation in high and middle latitudes, repeated pluvial phenomena in middle and low latitudes, and related worldwide fluctuations of sealevel. It is also marked by the appearance and intercontinental migrations of the modern horse, cattle, mammoths, camels, and man. It includes all of post-Pliocene time.[63] This definition is a compromise, but it is believed to be more readily usable and applicable over a much wider area than any other proposed.

This tentative definition takes account of the evidence of glaciation, pluvial stages, fluctuation of sealevel, and vertebrate fossils. It is defective in that it does not consider the evidence of invertebrate fossils. But, inasmuch as the separation of Pleistocene from Pliocene on this basis, as we have already noted, involves great difference of opinion, inclusion of the invertebrates in a definition is not at present justified.

The definition suggested eliminates the Recent as a formal stratigraphic subdivision. Kay and Leighton[64] disposed of the Recent in a different way by assigning to it the value of an age, immediately following the Mankato sub-age. Although not without advantages, this usage appears to be somewhat arbitrary in that it makes no provision for glacial subdivisions that in future may be recognized as important post-Mankato events. It seems better to use the term *recent,* like the term

[62] Daly 1929, p. 721.

[63] This definition was framed after consultation with Professor Carl O. Dunbar, Dr. Edwin H. Colbert, and Dr. H. G. Richards. It is not implied, however, that any of these authorities necessarily subscribe to this particular definition.

[64] Kay and Leighton 1933, p. 672.

postglacial, in an informal sense without exact stratigraphic definition.

The Pleistocene epoch is subdivided into *ages*, the major glacial and interglacial time units whose stratigraphic equivalents are *stages*.[65] It is true that *stage* as a stratigraphic term[66] is defined as a faunal zone, cutting across sedimentary facies, of smaller value than a group and of larger value than a formation. There appears to be no valid ground for objecting to the application of *stage* (and its time equivalent, *age*), in a somewhat different sense, to glacial features, for basically this application follows the original concept of *stage*. It establishes stratigraphic and time units that are of more than local significance and independent of local facies. In the older rocks this is made possible by the differences in successive faunas. In the Pleistocene, it is made possible by the effects of climatic changes. Even further, if, as seems likely, faunal zones in the Pleistocene marine sequence should prove to be controlled by climate, it should be possible to make direct correlation between glacial features and marine deposits without any break in stratigraphic terminology.

In North America the latest or Wisconsin Glacial age is subdivided into four *sub-ages* whose stratigraphic equivalents are *substages*. These units of the second order are differentiated on a basis of differences in degree of decomposition of drift sheets, conspicuous systems of end moraines, and fossil-bearing interglacial deposits of a minor order. It is probable that, as knowledge of northern North America grows, one or more additional sub-ages will be added to the four now recognized.

Knowledge of the pre-Wisconsin drifts is still too fragmentary to justify the recognition of sub-ages within any of the pre-Wisconsin glacial or interglacial ages.

SUMMARY

In summary, it is suggested that (1) the Tertiary period and Quaternary period be dropped from stratigraphic nomenclature, (2) the Pleistocene be considered as an epoch and be provisionally defined on a broad basis of glacial climates and fossil vertebrates and be considered (as defined by Lyell) to represent all of post-Pliocene time, (3) the term recent be used only in an informal sense, (4) the term postglacial be recognized as an informal term applicable only within geographically restricted areas. The problem of defining the Pleistocene is a large one and for its best solution should have the attention of a committee of specialists in the various fields involved. The present attempt is an effort to formulate a scheme that will be workable until such time as the matter is taken up by a fully qualified group.

[65] Cf. Eaton 1928; Kay 1931; Kay and Leighton 1933.

[66] First used by D'Orbigny in 1829.

On the basis suggested the classification would be as indicated in Tables 3 and 4.

TABLE 3. CLASSIFICATION OF THE POST-PLIOCENE, AND GLACIAL SUBDIVISIONS OF THE NORTH AMERICAN PLEISTOCENE

Era	Period (System)	Epoch (Series)	Age (Stage)	Sub-age (Substage)
Cenozoic	Cenozoic	Pleistocene	Wisconsin	Mankato
				Cary
				Tazewell
				Iowan
			Sangamon	
			Illinoian	
			Yarmouth	
			Kansan	
			Aftonian	
			Nebraskan	
		Pliocene		

(Interglacial units are shown in *italic* type.)

PLEISTOCENE STRATIGRAPHY OF CENTRAL UNITED STATES

On the basis of the characters enumerated above, there has been recognized in central United States a stratigraphic sequence consisting of four glacial stages and three interglacial stages. Within the Fourth Glacial stage, four glacial substages have been identified. This composite stratigraphic sequence has become a standard for the glacial deposits of the North American continent, and appropriately so, because the deposits concerned are continuous throughout wide areas and their stratigraphic relationships are for the most part well established. As it is very unlikely that a stratigraphic column as complete or as well exposed as this one

will ever be established in either the Atlantic or Cordilleran regions, where relief is much greater than in central United States, or the vast region of central and northern Canada, where glacial erosion predominated over deposition, the standard column already established is likely to become permanent. Probably its names will be applied more and more widely to stratigraphic units in other regions.

The fact that the glaciation of central North America had been multiple rather than single was recognized twenty years after multiple glaciation had been established in Europe. Orton and Winchell in 1873 and McGee in 1875 were the first to recognize interglacial deposits in central United States, and Hinde in 1878 recognized similar deposits at Toronto in Canada. All these deposits were identified on a basis of their fossil content. In 1878 T. C. Chamberlin differentiated two distinct drift sheets on strictly inorganic evidence.

In 1894 Chamberlin established the practice of naming the stratigraphic units after geographic localities, a practice that has continued through the present time. The development of the whole column and its present status are summarized in Table 4.[67] Although this table is believed to represent fairly the present trend of opinion, it does not represent universal usage. Until recently the Iowan drift was correlated by Leverett with the Illinoian and was regarded by others as an independent drift sheet having the value of a separate stage. Kay and Leighton[68] proposed pairing the Nebraskan and Aftonian ages into a single epoch, and the Kansan and Yarmouth, the Illinoian and Sangamon, and the Wisconsin and "Recent" into three additional epochs, all four epochs constituting the Pleistocene, which is given the value of a period. This grouping is artificial, introduces unnecessary names, and distorts the relative importance of the Pleistocene;[69] probably for these reasons it has not received strong support.

The Wisconsin drift began to be subdivided as early as 1899, when Leverett[70] recognized two distinct drifts, both of Wisconsin age, in northern Illinois. The Minooka and Marseilles moraines were considered as marking the outer limit of the Late Wisconsin drift. The Wisconsin drift older than these moraines was called the Early Wisconsin drift. In 1915 Leverett abandoned this classification on the ground that it introduced irreconcilable discrepancies in the implied shrinkage of the Lake Michigan and Lake Erie lobes of the Laurentide Ice Sheet.[71] Later

[67] Further details on individual units are given in U. S. Geol. Survey, Bull. 896, 1938.
[68] Kay and Leighton 1933.
[69] Cf. R. T. Chamberlin 1942, p. 918.
[70] Leverett 1899, p. 317.
[71] Leverett and Taylor 1915, p. 29.

TABLE 4. PRESENT STATUS OF THE GLACIAL

(Glacial units are shown in roman

Stage	*Substage*	*Named by*	*Original Reference*
Wisconsin		T. C. Chamberlin	J. Geikie, *Great ice age*, 3d Ed , 1894, p. 763.
	Mankato	M. M. Leighton	Sci., 77, 1933, p 168.
	Cary	M. M Leighton	Sci , 77, 1933, p. 168.
	Tazewell	M. M. Leighton	Sci., 77, 1933, p. 168.
	Peorian	F. Leverett	Jour. Geol., 6, 1898, p 246.
	Iowan	T. C. Chamberlin	Jour. Geol., 4, 1896, p. 874
Sangamon		F. Leverett	Jour. Geol., 6, 1898, p. 176.
Illinoian		F. Leverett	Jour. Geol , 4, 1896, p. 874.
Yarmouth		F. Leverett	Jour. Geol., 6, 1898, p. 239.
Kansan		T. C. Chamberlin	J. Geikie, *Great ice age*, 3d Ed., 1894, p. 755.
Aftonian		T. C. Chamberlin	Jour. Geol., 3, 1895, p. 270.
Nebraskan		B. Shimek	Geol. Soc. Am., Bull. 20, 1909, p. 408.

STRATIGRAPHIC COLUMN IN CENTRAL UNITED STATES

type; interglacial units in *italic* type.)

Type Locality or Region	*Remarks*
State of Wisconsin	Originally termed *East-Wisconsin*, its name was soon shortened to *Wisconsin* (Jour. Geol., 3, 1895, p. 270–277).
Mankato, Minnesota	Includes the Port Huron moraine and all younger drift. Post-Mankato substages are certain to be recognized as Canadian data are augmented.
Cary, Illinois	Includes the Cary moraine and younger drift up to but not including the Port Huron moraine.
Tazewell County, Illinois	Includes the Shelbyville moraine and younger drift up to but not including the Cary moraine.
Peoria, Illinois	When originally proposed, the Peorian had the value of an interglacial stage.
State of Iowa	Chamberlin had previously applied the names *East Iowan* (J Geikie, *Great ice age*, 3d Ed., 1894, p. 759) and *Iowan* (Jour Geol , 3, 1895, p. 270–277) to the drift sheet now called Kansan.
Sangamon County, Illinois	Includes Loveland formation (chiefly loess), named by Shimek (Geol. Soc Am , Bull. 20, 1909, p. 405), from Loveland, Iowa.
State of Illinois	When first proposed, the name was written *Illinois*, but this was soon changed to *Illinoian.*
Spoil of dug well, Yarmouth, Iowa	Since 1898 many exposures in southeastern Iowa have been identified.
State of Kansas	This name was applied in 1894 to the drift now called Nebraskan, but was soon shifted (Jour. Geol., 4, 1896, p. 872) to the drift sheet to which it is at present applied.
Afton, Iowa	First described by Bain (Iowa Acad. Sci., Proc., 5, 1898, pp. 93–98).
State of Nebraska	Unfortunately not well exposed in Nebraska. This drift has been called also *sub-Aftonian* and *pre-Kansan.*

Leverett adopted a threefold classification of the Wisconsin, consisting of Early, Middle, and Late substages.[72] This classification is less flexible than the one given in Table 3 and leaves no place for the later additions that are almost sure to be made when Canada has been more thoroughly studied.

A very complicated classification of the time from the end of the Pliocene to the present was offered by Antevs,[73] but it has not come into general use, probably because it would be very difficult to apply.

The name *Wabash beds* was proposed by Hay[74] for the deposits laid down "subsequent to the retreat of the ice sheet but prior to the present epoch." Not only is the name Wabash preoccupied, but also the definition is impractical, for the time of deglaciation varied greatly from place to place, and no basis has been found for subdividing "postglacial time."

In the next three chapters, the stratigraphic and geographic relations of the successive glaciations are set forth in greater detail.

[72] Leverett 1929*a*.
[73] Antevs 1931, pp. 2–4.
[74] Hay 1912, p. 13.

Chapter 12

GLACIATION OF NORTH AMERICA IN THE WISCONSIN AGE

INTRODUCTION

In describing the glacial and interglacial deposits of North America and in attempting to reconstruct from them the varied conditions of climate, drainage, and plant and animal ecology that prevailed during the Pleistocene epoch, we shall begin with the Wisconsin, the latest and youngest of the glacial ages, because we know far more about it than about the earlier Pleistocene times. Having in mind the reconstructed picture of Wisconsin conditions, very incomplete to be sure, yet coherent, we shall be in a better position to evaluate the evidence of pre-Wisconsin conditions than if we had considered Pleistocene events in strictly chronologic order.

Although at the height of Wisconsin time glacier ice was continuous across northern North America from the Atlantic to the Pacific, the former glaciers fall into two groups, according to their character and place of origin. The first group consists of the Cordilleran Glacier Complex, the network of former glaciers that occupied the mountains of western North America. The second group includes the Laurentide Ice Sheet, which spread over North America from Newfoundland to the Rocky Mountains, and the Greenland Ice Sheet which may have been continuous with it.

The descriptions of the extent and directions of movement of the former glaciers are based as far as possible on geologic evidence, which, however, is available in satisfactory strength only for parts of the Cordilleran region and for the southern peripheral part of the region covered by the former ice sheet. Therefore the description must be based in part on deduction from climatically reasonable assumptions. Until detailed geologic evidence has become available for the whole of northern North America some doubt must attach to the synthesis given here. But it is consistent with the evidence we do have, and its main outlines are believed to be sound.

CORDILLERAN GLACIER COMPLEX

General Distribution of Glaciers

So much of the northern Cordilleran region is known only in the most general way that it is not yet possible, except in a few intensively studied

localities, to discriminate between Wisconsin and pre-Wisconsin glaciation. There is direct evidence that one or more glaciations preceded the Wisconsin, and that in some places the Wisconsin ice was less extensive than the earlier ice. Nevertheless, in the discussion that follows it has been found necessary to treat the entire glaciated area as though it were a record of the Wisconsin glaciers. More detailed information will alter the picture, particularly in the eastern part of the mountain region, but it is unlikely to alter it radically.

The most conspicuous single element in the Cordilleran Glacier Complex as it existed during the Wisconsin maximum was the continuous and interconnecting mass of valley glaciers, piedmont glaciers, and an ice sheet — a mass that centered in British Columbia and stretched northwest to the Aleutian Islands and south to Mount Adams near the Columbia River, an overall distance of 2350 miles (Plate 3). The glaciers formed chiefly in the Coast Ranges and Cascade Mountains and to a lesser extent in the Rocky Mountains farther east. Most of the lower though still chiefly mountainous country between these two great chains was buried beneath ice that at first flowed inward from the two mountain chains. So completely did the piedmont glaciers fed from eastern and western ranges coalesce over this somewhat lower country, that the confluent mass has been sometimes called the "Cordilleran Ice Sheet," and indeed, if applied to central and northern British Columbia during the maximum of glaciation, this term seems fully appropriate. The area covered by coalescent glaciers is believed to amount to about 875,000 square miles.

On the Pacific coast proper the glaciers flowing down the western slope of the Coast Ranges or through the great transverse valleys such as the Fraser, the Skeena, and the Stikine, that drain the interior of British Columbia, calved off in the deep water immediately offshore. Where lowlands intervene between the mountains and the sea, as on the northern Pacific coast of Alaska and in the Puget Sound region in Washington, the valley glaciers coalesced into thick piedmonts before reaching deep water. On the eastern slope of the Rockies the Cordilleran glaciers met and coalesced with the Laurentide Ice Sheet. But in the interior, between the Rockies and the Coast Ranges, the ice found no ready escape and in consequence piled up and for a time constituted an ice sheet, the extent of which has not yet been determined. The width of this whole confluent mass from the west coast of Vancouver Island to Calgary, Alberta, beyond the eastern base of the Rockies, was 550 miles. Farther north, from Yakutat Bay in Alaska to the Mackenzie Mountains in the Northwest Territories of Canada, its width was very little less.

In addition to this vast confluent mass, the Cordilleran Glacier Complex included separate disconnected glacier groups and systems, each

centering on a mountain range or other highland mass. Several of these were in northern Alaska, the chief occupying the massive Brooks Range, 600 miles long and glaciated from end to end. Many others, embracing at least 75 separate areas, lay in western United States, the largest being the glacier plexus, 250 miles long, on the Sierra Nevada in eastern California. These glacier groups are described more fully elsewhere in this chapter.

These Cordilleran glaciers reached by far their most extensive development in British Columbia. They diminished both northward into Alaska and Yukon, and southward through western United States. Their distribution closely parallels the distribution of the relatively few and small glaciers of today. The greatest extent of glaciers in Wisconsin time, as now, was in regions (1) where the absolute amount of precipitation was large, (2) where a large proportion of the total precipitation fell as snow rather than as rain, and (3) where the summers were cool and short. In the Cordilleran region as a whole the combination of these conditions most favorable for building glaciers is found in coastal British Columbia. It becomes less favorable northward toward northern Alaska and Yukon owing to decrease in the first of these conditions; southward into the United States it becomes less favorable owing to decrease in the second and third.

Returning from an early exploring expedition in northwestern Canada, McConnell pointed out that existing glaciers are unknown in the Rocky Mountains north of about latitude 54°. At the Peel River portage, latitude 67°30′, he found that the winter snow had entirely disappeared before mid-July. Reasoning from such facts, he concluded that "climatic changes which would extend the present glaciers of the Bow and Saskatchewan far down their valleys might have little or no effect in imposing glacial conditions on this more northern region."[1]

Growth of the Glaciers

The growth of the Cordilleran glaciers took place concomitantly with a gradual descent of the regional snowline. The lowered snowline is recorded in a very general way by the altitudes of cirques abandoned by glaciers as the snowline rose once more.[2] This type of evidence is inexact not only as to the vertical position of the snowline but also as to the time when the snowline was lowered to a particular position. The snowline descended repeatedly during pre-Wisconsin glacial ages, and no doubt cirques formed during earlier glacial ages were reoccupied during later ones. Hence it is rarely possible to distinguish between the record of the

[1] McConnell 1890, p. 543.

[2] This matter is discussed more fully in Chapter 18.

Wisconsin snowline and that of earlier snowlines. However, as the limits reached by the Cordilleran valley glaciers during the several glacial ages did not differ among themselves vastly, it seems probable that all the snowlines referable to the maxima of the several glacial ages were somewhat alike in position.

It is clear that the agreement between the amount of precipitation and the extent of local glaciation was close. A glance at a glacial map[3] shows that former local glaciation in the Glacier National Park district, northwestern Montana, was more extensive than in the Rocky Mountain National Park district, Colorado, and that in the Mt. Rainier National Park district, Washington, it was still more extensive. Table 5 shows that the present-day relative precipitation on these three districts increases with increase in former glaciation. It shows further that the descent of the snowline during the glacial ages increased with increase in present-day precipitation. The same relationship has been found to exist in so many other parts of Cordilleran North America that it is probably

TABLE 5. COMPARISON OF PRESENT AND FORMER REGIONAL SNOWLINES IN THREE DISTRICTS IN THE CORDILLERAN REGION OF WESTERN UNITED STATES

	Mt. Rainier National Park, Washington	*Glacier National Park, Montana*	*Rocky Mountain National Park, Colorado*
Altitude of highest peaks (feet)	14,400	10,000	14,200
Mean annual precipitation (inches)	At Paradise (alt. 5500) 100 At Longmire (alt. 2800) 78	> 30	30
Altitude of present snowline (feet)	> 10,000	10,000	14,000
Altitude of low former snowline recorded by cirques (feet) (approx.)	4,500	5,500	10,000
Difference between the two snowlines (feet)	> 5,500	> 4,500	4,000

universal throughout this region. This relationship argues strongly for the belief that throughout the Pleistocene epoch the climatic belts have borne much the same relations to each other as they do at present, that although they have expanded and contracted and shifted somewhat in latitude with the passage of time their relative positions have remained much the same. It argues also for the belief that should a general cooling

[3] Cf. Flint and others 1945.

now occur, with all other factors remaining unchanged, the Cordilleran glaciers would once more develop as they did during the onset of the Wisconsin Glacial age.

Over much of the interior of British Columbia evidence of erosion on the higher ridges is so slight that for a time it was thought that these ridges had never been covered by ice. After many years of study Johnston concluded that at the maximum of at least one glacial age this region had been buried beneath an ice sheet at least 3000 feet thick and having very little radial flow, probably because it was firmly confined between two high mountain barriers. He found that evidence of inward flow of the former glaciers from the mountains toward the interior is much stronger than evidence of outward flow from the interior toward the mountains. From this he reasoned that local valley and piedmont glaciers had occupied this region through a far longer time in the aggregate than had the ice sheet, which was the result of temporary conditions peculiar to the glacial maximum.[4]

In northwestern British Columbia Kerr found the glacial history to be much the same. In the region of the Stikine and Taku valleys, he recognized four successive phases of glacier growth:[5]

1. An *alpine* phase, in which small- and medium-sized valley glaciers developed on the higher mountains. Deglaciation has brought the region back to this condition at present.
2. An "intense-alpine" phase, in which growth and coalescence of the valley glaciers led to the development of long and massive trunk valley glaciers that filled the main valleys transecting the Coast Ranges.
3. A "mountain-ice-sheet" phase, in which further growth developed extensive piedmont glaciers along both flanks of the Coast Ranges; the ice had become so thick that it virtually covered the Coast Ranges.
4. A "continental-ice-sheet" phase, marked by glacier flow westward across and independent of the Coast Ranges.

A similar development was inferred in southern British Columbia.[6] The transition from 3 to 4, as well as the evidence of reversal of direction of flow cited by Johnston, implies that the ice divide, the center or axis of radial outflow, shifted from the crestlines of the mountains themselves, inward to positions over lower land. From the crest of the Coast Ranges the shift was eastward, but from the crest of the Rockies it was westward.[7] When the two axes of radial flow coalesced, the Cordilleran

[4] Johnston 1926, p. 137.
[5] Kerr 1936, p. 681.
[6] N. F. G. Davis and Mathews 1944.
[7] Enquist (1916) deduced an eastward shift from the crest of the Rockies, and in this opinion was followed by Antevs (cf. 1945). These and other opinions are discussed in Flint 1943, p. 335.

glacier complex temporarily became (or rather centered in) an ice sheet.

The shift in the ice divide on the Coast Ranges could have occurred in this way (Fig. 49): The chief locus of snowfall was on the western side of the mountains that faced the moist maritime air masses moving against them from the Pacific Ocean, just as it is today. Although the total accumulation of snow and therefore glacier ice probably was greater on the western side than on the eastern, nevertheless the removal of ice by glacier flow was far greater on the western side. This is because the average western slope is steeper and, more important, because on the

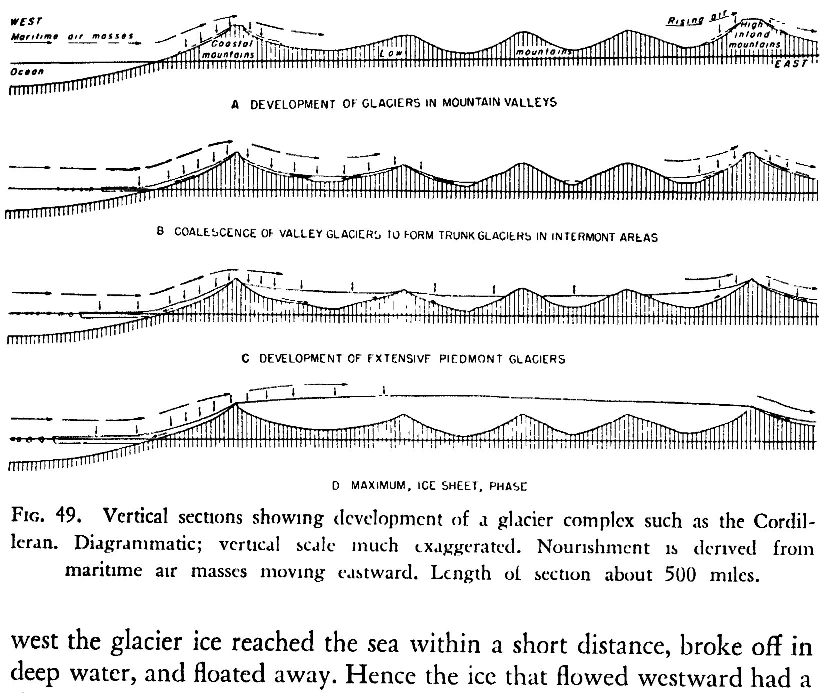


FIG. 49. Vertical sections showing development of a glacier complex such as the Cordilleran. Diagrammatic; vertical scale much exaggerated. Nourishment is derived from maritime air masses moving eastward. Length of section about 500 miles.

west the glacier ice reached the sea within a short distance, broke off in deep water, and floated away. Hence the ice that flowed westward had a short and swift journey on a steep slope. But the ice that flowed eastward had a slower journey on a longer and much gentler slope and was not subject to calving at its termini. It may have been favored to some extent also by wind-drifting of snow toward the east. Thus, despite probably greater accumulation on its western side, the ice there drained away as fast as it accumulated, whereas on its eastern side, despite accumulation that may have been smaller, drainage was slower and terminal losses were less. In consequence the east-flowing valley glaciers waxed and coalesced in the rough "plateau" country east of the Coast Ranges into a vast piedmont apron that spread out, thickened, and, with continued

accumulation, "backed up" into the valleys until its summit covered much of the mountain crest. The ice on the east had now become a true ice sheet, but the ice west of the mountain crest still consisted primarily of valley glaciers. A greater *net* accumulation, whether by direct snowfall or wind-drifting or both, east of the mountain crest than west of it, in time would have brought about an eastward shift in the center of radial flow despite a probably greater total precipitation on the west side.

Relation to the Laurentide Ice Sheet

Meanwhile the glaciers developing on the Rocky Mountains 300 miles to the east were evolving in a somewhat different way. The snowfall on the Rockies must have been considerably less than that on the Coast Ranges, and at any one time the glaciers must have been correspondingly smaller. There was no sea water to bring about extensive calving; the valley glaciers that flowed down the slopes of the Rockies coalesced to form piedmonts on both flanks of these mountains. At length, while the eastern piedmonts met and coalesced with the Laurentide Ice Sheet that was encroaching toward the Rockies from the east,[8] the western piedmonts met and coalesced with the thick ice flowing eastward from the region of the Coast Ranges. From this latter coalescence ultimately grew the ice-sheet phase of the Cordilleran Glacier Complex. In northern British Columbia the ice at this time is estimated to have been 10,000 feet thick. The direction of flow in its western part was reversed, for the evidence indicates that ice from the interior of British Columbia flowed westward across the Coast Ranges. Whether the snowfall was sufficiently great and the resulting flow sufficiently strong to bring about eastward flow from the interior eastward across the Rockies is a matter of doubt. The western part of the complex must have been more vigorous and active than the eastern at all times.

In northern Montana Cordilleran ice reached at most 40 miles east of the main mountain front. Farther north, in Alberta, the ice reached Calgary, 50 miles east of the mountains, and in the latitude of Edmonton it reached slightly farther. Hence it appears that the piedmont ice from the Rocky Mountains did not extend far east on to the plains. Probably this was true even before this ice met and coalesced with the great ice sheet flowing from the east. The Coast Ranges and the Rockies and the highlands between them produce a strong rain shadow. Glaciers on the east slope of the Rockies were thereby prevented from receiving abun-

[8] Along the Alaska Highway at latitude 58°40′ granitic boulders derived from the Canadian Shield to the east occur within 5 miles of the Rocky Mountains front and up to an altitude of 4000 feet.

dant nourishment. The westerly winds yielded snow to the much more massive glaciers farther west and then, as they descended the east slope, were adiabatically warmed and precipitated little.[9] The growth of the Cordilleran glaciers therefore was limited by the snow accumulation in the vicinity of the mountains themselves. In this respect these glaciers differed sharply from the Laurentide Ice Sheet, which soon spread away from the highlands in which it originated.

Conditions in the Far North

Toward the north the vast reservoir of Cordilleran ice was virtually dammed up and pooled in the corridor, 300 miles wide and more than 1000 miles long, between Coast Ranges and Rockies. From this pool it sought escape down the broad Yukon drainage basin south and east of the vicinity of Dawson.

The difference between the strong glaciation of the Coast Ranges and the less strong glaciation of the mountains farther inland and to leeward becomes rapidly more pronounced north of latitude 60°. Ketchikan, on the coast at latitude 56°, has 150 inches of precipitation annually. Inland from the Coast Ranges, nonglaciated summits rise above glaciated lower lands, the upper limit of glaciation declines northwestward, and finally, in the interior, evidence of glaciation ceases. This progressive change results in part from increasing distance from the paths of moist air masses that move eastward from the Pacific, and in part from the strong rain-shadow effect of the massive coastal mountains in Alaska, which are higher than the mountains farther south. Mt. McKinley, the highest, reaches 20,300 feet.

The Coast Ranges are glaciated right out through the Aleutian chain of islands, all of which (except the low, nonmountainous islands) appear to have been glaciated. Inland the most extensive glaciers formed on the massive Brooks Range in northern Alaska. But even these were thin and only partly coalescent, and they failed to cover the Arctic coastal lowland at their northern base. The explanation can only be deficient precipitation. Present-day figures show only 6 to 10 inches annually at the northern base of the range.

Between the Coast Ranges and the Brooks Range the isolated mountain groups standing above this broad region of hills and lowlands carried thin glaciers.

[9] However, as long as the margin of the Laurentide Ice Sheet was in contact with the Rocky Mountains glaciers the rain shadow must have been less effective than it was both before and after the coalescent condition existed.

Sea Ice in the Pacific-Arctic Region

With climates as they are at present, floating sea ice (the greater part of which is frozen seawater) occupies most of the Arctic Sea and extends southward into the North Atlantic region and into the Bering Sea. Pack ice, that is to say continuous or nearly continuous sea ice, extends down to nearly latitude 58° in the Bering Sea during the winter, and single ice masses broken from the main pack extend much farther (Plate 2). In summer the southern edge of the pack recedes north of Bering Strait.

In the absence of direct evidence of conditions at sea during the maximum of any glacial age, we can do no more than make a reasonable guess at the extent of sea ice at such times. There can be little doubt that the termini of many glaciers (including large piedmont glaciers) debouching into the Pacific from the Coast Ranges of Alaska and British Columbia were afloat, like the Ross Shelf Ice in Antarctica today. It is not likely, however that they extended far out to sea. The Ross Shelf Ice ends in very cold water, but the Pacific coast of North America is bathed in the comparatively warm water of the North Pacific Current, which should have been able to melt the "barrier" or terminus of the shelf ice effectively. Icebergs detached from the floating glaciers would have drifted westward, with the prevailing current, toward the Aleutian Chain. It seems unlikely that any ice would have drifted southward toward California.

The comparative warmth of the North Pacific Current would likewise have inhibited the formation of sea ice in the North Pacific Region and also would have prevented its leaking southward from the Bering Sea through the many straits in the Aleutian Chain. It seems probable that the cold Bering Sea would have been filled with pack ice, which would have reached south to the Aleutians but not past them into the warmer water of the Pacific.

Along the Bering and Arctic coasts of Alaska the glaciers did not reach the sea; hence there would have been no bergs from those sources. But throughout the glacial maxima the surfaces of these seas must have been frozen, with but few openings.

Southern Limit of Continuous Ice

Toward the south coalescent glaciers occupied the Cascade Mountains uninterruptedly to the Columbia River. East of these mountains the margin of the continuous ice crept across the 49th parallel and reached nearly a hundred miles farther into the State of Washington, where it attained its farthest extent along the northern margin of the hot, dry

Columbia Plateau. Evidence as to ice thickness and as to direction of flow gathered from this region suggests that here the glacier ice at its maximum approached though it never fully attained the status of an ice sheet.[10] The upper surface of the ice appears to have been concave-up, its outer margin was more extended near both chains of mountains than in the lower country between them, and the directions of the glacial striae show a distinct component inward from the Coast Ranges (and Cascade Mountains) on the west as well as from the Rocky Mountains on the east. Evidently in this latitude the net accumulation of snow on the Coast Ranges was insufficient to shift the ice divide to the east, and in consequence the ice in the interior seems not to have grown beyond a much-thickened piedmont condition, essentially the "mountain-ice-sheet" phase of Kerr. South of the 49th parallel the glaciers between the ranges diminished in thickness and disappeared, reflecting the southward-rising regional snowline.

The west flank of the Cascade Mountains was covered by thicker and more nearly continuous ice than the drier east flank. North of latitude 47° the ice on the west flank was very nearly continuous, merging westward with the massive Puget lobe, a piedmont glacier 2500 feet thick fed by ice flowing from the highland region immediately to the north.

The Strait of Georgia was filled with piedmont ice fed from the mountains on the mainland to the east and on Vancouver Island to the west. This ice flowed south and then west through the Strait of Juan de Fuca forming what has been called the Juan de Fuca piedmont lobe. It seems likely that this ice ended in a floating shelf, which may have fringed the entire west coast of Vancouver Island.

The phases of growth of the Cordilleran complex parallel closely the probable development of the Greenland Ice Sheet, which, however, appears to have expanded to a status of more powerful radial outflow on its eastern flank than did the Cordilleran mass. This difference may have resulted from greater nourishment of the Greenland ice by virtue of its being fringed by the sea on both its sides instead of on its western side alone. The Scandinavian Ice Sheet, too, grew up in a closely similar way, as is set forth in Chapter 15.

Highland Glaciers beyond the Continuous Ice

United States

Within western United States south of the limit of continuous ice there were at least seventy-five separate areas of glaciers having a combined area of roughly 35,000 square miles. Each area centers in a high-

[10] Flint 1937, p. 226.

land — a mountain range, group of ranges, or plateau. Most of the areas were occupied by valley glaciers, singly or in groups, but some were complex, including plateau and mountain ice caps. The distribution of these areas is indicated on the map, Plate 5, which shows the extent of glaciation regardless of stage; hence all the areas are larger, and some of them are considerably larger, than they would be if they represented only the Wisconsin stage.

Furthermore the areas are necessarily generalized. In consequence one area may include within it smaller areas that were never ice covered. This is notably true of the Klamath Mountains area (No. 7), which embraces many small glaciated areas, not all of them coalescent. It is true also of the Salmon River Mountains area (No. 14) which, like the Klamath Mountains area, is not yet well enough known to permit more detailed mapping.

From the map it is apparent that the largest individual area of former glaciation lies in what we may call the Yellowstone-Teton-Wind River highlands (No. 45), which include many individual mountain ranges and plateaus. Each of the higher ranges supported numerous valley glaciers, which flowed out and in some places coalesced as piedmont glaciers on the lower lands beyond. The Yellowstone Plateau was overwhelmed by coalescent piedmont ice which seems to have thickened and developed radial outflow from the plateau itself. The Beartooth Plateau to the northeast had an ice cap of its own.

The second largest individual area of former glaciation is in the Sierra Nevada (No. 8). In the southern part of the range the ice took the form of valley glaciers that were coalescent only near their heads, along the high axis of the range. Toward the north, where the mountain crest is higher, the glaciers increased in length and thickness and coalesced to form an ice cap that covered the intervalley ridges and above which only a few high peaks projected as nunataks. Still farther north, where altitudes diminish, the ice cap once more gave way to a complex of valley glaciers. Where the ice cap existed the ice divide was for a time displaced somewhat toward the west from the position of the present drainage divide. From the ice divide, glacier flow occurred eastward *up* the valley heads and across the crest of the range. However, east of the crest the ice never had any great extent, apparently because of aridity induced by the great height of the mountains.[11]

The former distribution of glaciers on the highlands of western United States is related principally to altitude, to latitude, and to the principal source of moisture — Pacific air masses. The altitude factor, which largely determined the amount of orographic precipitation on

[11] Matthes 1930; F. E. Matthes, *unpublished.*

TABLE 6. PRELIMINARY DATA ON GLACIATED AREAS IN WESTERN UNITED STATES
(This list is incomplete, additional glaciated areas undoubtedly exist in this region.)

Reference Number on Map, Plate 5	*Name of Highland*	*Highest Known Altitude of Highland (feet)*	*Type of Glacier*	*Number of Stages or Substages Recognized*	*General Altitude of Lower Cirque Floors (feet)*	*Lowest Known Altitude Reached by Glaciers (feet)*	*Reference (selected; no attempt is made to give all references)*
1	Juan de Fuca piedmont lobe		piedmont	2		0	Johnston 1923, pp. 39–54.
2	Puget piedmont lobe		piedmont	2		0	Bretz 1913, Hansen and Mackin 1940.
3	Olympic Mountains	7,954	ice cap?: valley glaciers		4500–5000	0	
4	Cascade Mountains in Washington	14,415	complex	3	4500–5000 in N 5000 in S	0	I. C. Russell 1900; Page 1939.
5a–c	Cascade Mountains in Oregon						Scattered.
a	Mt. Hood district	11,225	complex	2	4500–5000	200	
b	Mt. Jefferson district	10,500	complex	3	5500		Thayer 1939.
c	Crater Lake district	8,938	complex	3	6000		
6	Mt. Shasta	14,161	complex	Obscured by volcanism		3900	Diller 1895.
7	Klamath Mountains (many separate areas)	9,345	valley glaciers	3?	6500	2500	Hershey 1900, 1903.
8a–c	Sierra Nevada						
a	Lassen Peak district	10,457	complex	Obscured by volcanism	7000–7500		Howel Williams 1932, p. 381.
b	Mt. Elwell district	8,615	valley glaciers		7500		Diller 1908, p. 80.
c	Principal area	14,495	complex, including ice sheet in northern part	4	7500 in N 9000 at Lat. 38° 11,500 in S	2000 (W. flank) 4700 (E. flank)	Blackwelder 1931; Matthes 1930; 1933, Knopf 1918, pp. 92–105
9	San Bernardino Mountains	11,485	valley glaciers		10,500	8500	Vaughan 1922.
10a, b	Okanogan and Spokane lobes		piedmont or ice sheet	2+?		1000	Flint 1937.
11	Mountains in N. W. Montana and N. Idaho	10,438	complex	2+?	5500 in N 7000 in S	2000	Anderson 1927; Alden in Flint and others 1945.
12a–c	Bitterroot Mountains	10,131	complex	3	6000 at Lat. 47°30′ 6500 at Lat. 46°		Lindgren 1904, p. 51, Jeanne Russell 1926.
13a–c	Clearwater Mountains	8,150	complex		6500 in NE 7000–7500 in SW		Lindgren 1904, p. 67.
14a, b	Salmon River Mountains	12,078	complex	2–3	8500 in N 9000–9500 in SW	4000	Capps 1940, Ross 1938.

15	Lookout Peak district	8,124	valley glaciers		7500		
16	Seven Devils Mountains	9,387	valley or cirque glaciers				
17	Wallowa Mountains	10,037	valley glaciers and ice cap	2 Wisconsin substages	7500	4000	W. D. Smith and others 1941.
18	Elkhorn Mountains	8,922	valley glaciers		7500		
19	Strawberry Mountains	9,600	cirque glaciers?				
20	Owyhee Mountains		cirque glaciers?				
21	Steens Mountain	9,354	cirque and valley glaciers			9000	W D Smith 1927
22	Santa Rosa Mountains		cirque glaciers				
23	Jarbidge Mountains and adjacent ranges	9,911	cirque glaciers				Blackwelder 1934
24	Independence Range	11,000	cirque glaciers	2 Wisconsin substages			Blackwelder 1934.
25	Pilot Range	10,704	cirque glacier				Blackwelder 1934
26a, b	Ruby-East Humboldt Mts.	11,356	valley glaciers	2 Wisconsin substages	9000–9500	6100	Sharp 1938.
27	Shoshone Mountains	10,322	cirque glacier	1			Blackwelder 1934.
28a, b	Toiyabe Range	11,775	cirque glaciers				Blackwelder 1934.
29	Wassuk Range	11,303	cirque glacier	1			Blackwelder 1934.
30	Sweetwater Range	11,646	valley glaciers	2+?		9300	Blackwelder 1934.
31	White (Inyo) Mountains	14,242	valley glaciers	2 Wisconsin substages	12,000–12,500	6500	Blackwelder 1934.
32	Spring Mountain	11,910	cirque glacier			9300*	C. R. Longwell (unpubl.)
33	Snake Range	13,047	cirque glaciers				
34	Deep Creek Mountains	12,101	valley glaciers	2 Wisconsin substages			Blackwelder 1934.
35	Tushar Mountains	12,173	valley glaciers		10,500+		
36	Glenwood Mountain	11,223	valley glaciers or plateau ice cap				
37	San Francisco Mountains	12,794	valley glaciers	2 Wisconsin substages +1 older (possibly obscured by volcanism)	11,000–11,300	8500	Sharp 1942c.
38	Aquarius Plateau	11,000	plateau ice cap	Wisconsin	11,000		Gould 1939.
39a, b	Fish Lake Plateau	11,500	plateau ice cap?				Dutton 1880.
40	Wasatch Plateau	11 000	valley glaciers	Wisconsin	10,000–10,500	7800	Spieker and Billings 1940.
41	Stansbury Range	11,031	valley glaciers	2 Wisconsin substages		8500	Blackwelder 1934.
42	Oquirrh Range	10,585	valley glaciers	2 Wisconsin substages			Blackwelder 1934.

* Anomalously low because of extraordinary local topographic conditions.

TABLE 6. PRELIMINARY DATA ON GLACIATED AREAS IN WESTERN UNITED STATES (*Continued*)
(This list is incomplete; additional glaciated areas undoubtedly exist in this region)

Reference Number on Map, Plate 5	*Name of Highland*	*Highest Known Altitude of Highland (feet)*	*Type of Glacier*	*Number of Stages or Substages Recognized*	*General Altitude of Lower Cirque Floors (feet)*	*Lowest Known Altitude Reached by Glaciers (feet)*	*Reference (selected; no attempt is made to give all references)*
43a-d	Wasatch Mountains	12,008	valley glaciers	3	8500–9000 in N 10,000 in S	5000	Atwood 1909.
44	Uinta Mountains	13,500	complex	3	9500–10,500	6600	Atwood 1909, Bradley 1936.
45a-f	Yellowstone-Teton-Wind River highlands		complex				
a	Madison and Gallatin ranges	10,000			9000		
b	Snowy Range and Beartooth Plateau	12,850	plateau ice cap; valley glaciers	2–3	10,000		Bevan 1944.
c	Teton Mountains	10,685	valley glaciers	3	9000–9500		Horberg 1938, p. 72.
d	Absaroka Range	12,202	complex		10,500		Weed 1893; Horberg 1940.
e	Wyoming and Salt Creek ranges	10,763	valley glaciers		9000–9500		
f	Wind River Range	13,785	complex	3	10,500		Blackwelder 1915, p. 321.
46	Lost River Range	12,655	valley glaciers		9500–10,000		
47	Lemhi Range	11,025	valley glaciers		9500–10,000		
48a, b	Beaverhead Mountains	10,960	valley glaciers				
49	Centennial Mountains	10,211	valley glaciers				
50	Snowcrest Range	10,546	valley glaciers				
51	Tobacco Root Mountains	10,267	valley glaciers				
52	Pioneer Mountains	9,210	valley glaciers				
53	Red Mountain	10,150	valley glaciers				
54	Anaconda-Flint Creek-Sapphire Mountains	10,475	valley glaciers		7000–7500		
55	Deer Lodge Mountains	8,789	valley glaciers				
56	Crow Creek Mountains	9,432	valley glaciers		7500–8000	6300	Weed 1902.
57	Big Belt Mountains	9,478	valley glaciers				
58	Little Belt Mountains	9,000	valley glaciers				Alden 1932.
59	Castle Mountains	8,606	valley glaciers	2	8000	5000	Alden 1932, Weed and Pirsson 1896.

60	Crazy Mountains	11,214	valley glaciers	1	8700–9200		Alden 1932, G. R Mansfield 1908.
61	Big Horn Mountains	13,165	valley glaciers	3?	10,500–11,000	6200	Darton and others 1906*a*, 1906*b*
62	Medicine Bow Mountains	12,005	valley glaciers	Wisconsin substages +?	10,000	7500	W. W. Atwood, Jr, 1937; Ray 1940, p. 1874.
63	Park Range	12,220	valley glaciers		10,500		
64	White River Plateau	12,001	valley glaciers		11,000		
65a-c	Colorado Rockies		complex				
a	Ranges with coalescent glaciers	14,255	complex	Wisconsin substages +?	11,000–11,500 at N 12,000 at S	8000	Ray 1940; Capps and Leffingwell 1904, Bastin and Hill 1917, p. 57; Ives 1938, Lovering 1935.
b	Mount Evans district	14,260	valley glaciers	Wisconsin substages	12,000	< 9000	Ray 1940, p. 1877
c	Pikes Peak district	14,110	valley glaciers	Wisconsin substages		10,000	Ray 1940, p. 1880.
66	Grand Mesa	11,500	plateau ice cap?	2?		11,500	Henderson 1923.
67	La Sal Mountains	13,089	cirque glaciers	1	9500–10,000		Gould 1927.
68a-c	Sangre de Cristo Mountains						
a	(Northern) Sangre de Cristo Mountains	13 301	valley glaciers		11,000 in W 10,500 in E		Ray 1940.
b	Blanca Peak district	14,390	valley glaciers	Wisconsin substages		8800	Ray 1940, p. 1882.
c	Culebra and other ranges	13,546	valley glaciers	Wisconsin substages	11,000–11,500	8500	Ray 1940, pp. 1887–1901.
69	West Spanish Peak	13,623	valley glacier	Wisconsin substages	11,000	9000	Ray 1940, p. 1885.
70a	San Juan Mountains	13,725	ice caps and valley glaciers	3	12,000		Atwood and Mather 1932.
70b	Canjilon Divide	10,000	[extension of San Juan Mountains glaciers]	1 (early)			H. T. U. Smith 1936.
71	Sierra Blanca	12,003	cirque glacier	Wisconsin			H T. U. Smith and Ray 1941.

highland areas, is illustrated by the mountain ranges in the southern part of the Basin-and-Range region, lying between the Sierra Nevada on the west and the Rocky Mountains on the east. There only ranges reaching altitudes of more than 11,000 feet have yielded evidence of glaciation. The many ranges of slightly lesser altitude appear not to have carried ice, presumably because below the critical altitude snowfall was too slight and ablation too great. The latitude factor is shown by the continuously decreasing altitudes of glaciated highlands from south to north. The factor involving moisture source is revealed by the intense, low-level glaciation near the Pacific coast compared with the less intense, higher-level glaciation in the Rocky Mountains situated well inland.

Nearly all the glaciated mountain areas are marked by cirques, most of them well developed. It was pointed out in Chapter 6 that the general level of cirque excavation — the general altitude of the floors of the cirques — is a rough indication of the altitude of the regional snowline at the time the cirques were made. Where the glaciers were small the cirque floors may approximate the regional snowline rather closely; but, where snow was more abundant and glaciers expanded and descended to lower altitudes, the snowline descended also, reaching down well below the cirques. Consequently, in order to approximate the former regional snowline, a variable factor must be subtracted from the general level of cirque excavation, depending on the bulk and extent of the former glaciers.[12] As this factor is somewhat uncertain, it seems best to show the general level of cirque excavation, a feature capable of being measured directly, rather than the inferred regional snowline.

This is shown in Table 6 and by form lines in Plate 5. The data are generalized, because cirque floors are as much as 500 feet higher on the western and southern flanks of highlands than on the eastern and northern flanks. Also, on some highlands cirque floors are found throughout a considerable range of altitudes, resulting in part from local variations in lithology and probably in part from fluctuation of the regional snowline during a glacial age. In such highlands the altitudes of the lower rather than the higher cirques have been selected, as probably representing excavation during the maxima of the glacial ages. As already indicated it seems probable that the same cirques were occupied repeatedly during successive glacial ages, and that therefore no correlation can be made between a particular set of cirques and a particular glacial age. However, future refinements of study may show that some correlation is possible.

[12] In many of these mountains today the climatic snowline has risen, with secular increase in temperature, to positions above the cirque floors and even above the highest mountain summits.

At any rate the form lines drawn in Plate 5 clearly show a northward and seaward descent of the general level of cirque excavation. The regional snowline at a glacial maximum would have the same trend, but its slope would be steeper and it would reach lower altitudes on the west and north.

Mexico

Ixtaccihuatl (17,425 ft.), a volcanic cone on the plateau of Mexico, today has glaciers that reach down to about 15,000 feet. During at least one of the glacial ages glaciers descended to nearly 11,000 feet.[13] Two neighboring volcanoes, Orizaba (18,650 ft.) and Popocatepetl (17,950 ft.), are still higher and probably would have been ice-clad if they were as high during any glacial maximum as they are today. Orizaba is in a solfataric phase, but Popocatepetl is still active; hence neither cone now has glaciers. Both are undissected. It is not known whether former glacial features have been destroyed by volcanism or whether the cones were not built up above the regional snowline until very recently.

THE LAURENTIDE ICE SHEET

The Glaciated Region

The region covered by the Laurentide Ice Sheet, amounting at the Wisconsin maximum to about 4.8 million square miles, is shown in Plate 3. The southern border of this region is clearly defined and is well known. The western border is a narrow zone in which drift derived from the east is mingled with drift brought by Cordilleran glaciers from the mountains on the west; its position is known at only a few places. The northern border lies beneath the Arctic Sea and in part perhaps across some of the most northerly of the Arctic islands, on which we have virtually no glacial information. As the eastern border lies to seaward of the present coast, the extent of the glaciated region beyond the coast is unknown, though southward from Newfoundland the broad continental shelf, now only slightly submerged and receiving much precipitation, probably was glaciated extensively. Both the character of the sea-floor sediments and the forms of submerged troughs support this deduction.[14]

Directions of Flow

In this region the directions in which the glacier ice flowed are inferred from end moraines, striations, indicator stones, drumlins, eskers, and the asymmetric glacial abrasion of hills. These features are known

[13] Jaeger 1925, p. 371.
[14] Cf. Shepard 1931; 1932.

in detail only in the southern peripheral belt. There, and in the less-well-known eastern, western, and northern peripheral belts, they combine to record flow in radially outward directions. Prior to the 1880's, when only the southern belt was known, it was vaguely supposed that the ice sheet had spread southward from the north polar region. But when evidence of northward spreading of the glacier was found in Arctic North America it became apparent that the ice sheet had originated in much lower latitudes. It was named the *Laurentide Ice Sheet.*[15]

Explorations in the little-known country both east and west of Hudson Bay gradually revealed the existence of striations pointing more or less radially outward from several ill-defined areas in these regions. For some time thereafter it was generally believed that these areas were the "centers" from which the ice spread out, and the names *Labradorian* and *Keewatin* were given to two supposed distinct ice sheets thought to have spread from them. Later it became apparent that the striations from which these "centers" are inferred must have been made during a very late phase in the history of the glacier and have little overall significance. Each center of radial flow implies the former existence, not necessarily of a distinct glacier, but of a domelike part of the ice sheet from which the ice spread outward. Some centers, like those in eastern Quebec, western Newfoundland, and New Brunswick, coincide with highlands and were localized by exceptional snowfall induced by topography. Others, like those clustered around Hudson Bay, are situated on comparatively low land and must have been localized by irregularities in the distribution of snow accumulation on the ice sheet itself. Undoubtedly the centers on record existed during a late phase in the deglaciation process. Earlier centers there must have been, but the striations that recorded them were largely rubbed out by later flow in other directions, as the area and thickness of the ice sheet changed.

From striations, end moraines, drumlins, and eskers, because of their late time of origin, very little has been learned about the manner in which the ice sheet spread from its source areas to reach its greatest size. From indicator stones it is apparent that the glacier expanded radially outward through the western, southern, and eastern parts of the region it ultimately covered (Fig. 30). From end moraines and other features the earlier phases of its shrinkage are easily traced. As might be expected, early shrinkage was roughly concentric.

Because it is related to a highland well situated to receive abundant snowfall, the Labradorian center (lying more in Quebec than in Labrador) is believed to have been the area of origin of a separate ice sheet

[15] The history of the study of the ice sheet and the names ("Labradorian"; "Keewatin") by which it has been known are outlined in Flint 1943.

which by coalescence with glaciers flowing outward from other highlands in eastern North America formed the Laurentide Ice Sheet. However, no compelling evidence has yet been brought forward to indicate that a separate "Keewatin Ice Sheet" ever existed. Properly the term Keewatin refers only to a center or group of centers of radial outflow of ice, that existed late in the Wisconsin age.

Growth of the Ice Sheet

The Laurentide Ice Sheet was immigrant into most of the region it covered. It is probable that it originated as small valley glaciers in highlands in the northeastern part of the continent; that these glaciers expanded into piedmont glaciers on the lower lands to the west and south; that the piedmonts thickened and coalesced to form an ice sheet; and that, nourished by moisture-bearing winds from the south and west, the ice sheet expanded until it covered nearly 5,000,000 square miles.[16]

The highlands in which the ice originated are (1) the mountains of Baffin and Ellesmere islands, (2) the mountains of coastal Labrador, and (3) the highlands of eastern Quebec. The mountains on Baffin Island are 500 miles long and reach extreme altitudes of 10,000 feet, whereas on Ellesmere Island the peaks locally reach 13,000 feet.

The Labrador mountains extend from latitude 60° down to about 53°. Although not so high (4300 feet to more than 5000 feet) as the islands to the north, they receive greater precipitation. The highlands of eastern Quebec lie in the central part of the peninsula bounded by Hudson Bay, Hudson Strait, the Labrador Sea, and the St. Lawrence. They form an irregular mass some 150 miles in diameter, their highest parts are more than 3000 feet above sealevel, and they constitute a region of abundant precipitation. Both Baffin and Ellesmere islands have plateau ice caps today; there are no glaciers in Quebec and only a few insignificant ones in Labrador. Mean annual precipitation ranges from 10 inches to 30 inches near sealevel; high in the mountains it is probably considerably greater. Winters are cold, and summers are cool and short. The prevailing winds reach the highland region from the southwest and west, bringing moist maritime air from the Atlantic, the Gulf of Mexico, and the Pacific, and to a far smaller extent from the Arctic Sea as well.

If the temperature were to fall by a very small amount the glaciers on Baffin and Ellesmere islands would increase and new glaciers would form in the Labrador mountains and in the highlands of eastern Quebec. Following the reasoning set forth in Chapter 22, we shall assume that it

[16] This theory of origin is detailed in Flint 1943; see also Flint and Dorsey 1945*b*.

was through a reduction of temperature that the growth of glaciers in northeastern North America in the Wisconsin age began.[17]

As they expanded, the valley glaciers flowing down the eastern slopes of these highlands calved into Baffin Bay, Davis Strait, and the Labrador Sea. On the western slopes the glaciers coalesced to form piedmont glaciers on the lowlands beyond, for at that time Hudson Bay and the Foxe Basin probably did not exist.[18] The expanding piedmont ice began to compete with the mountains back of it for a share of the available snowfall. As the ice thickened and spread, chilling the air above and beyond it and creating a new topographic barrier to atmospheric movement, it received increasing snowfall at the expense of the mountains. As it

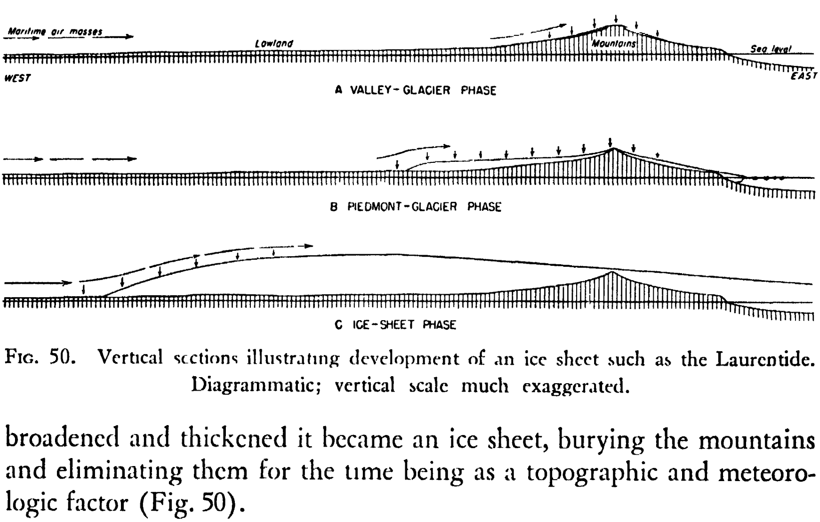


FIG. 50. Vertical sections illustrating development of an ice sheet such as the Laurentide. Diagrammatic; vertical scale much exaggerated.

broadened and thickened it became an ice sheet, burying the mountains and eliminating them for the time being as a topographic and meteorologic factor (Fig. 50).

Probably before this time an ice sheet had begun to form in the highlands of eastern Quebec and had merged with the Labrador ice. Spreading northward, the combined mass met and coalesced near Hudson Strait with the ice from Baffin Island and other islands to the north. With this event the Laurentide Ice Sheet proper could be said to have come into existence.

The growing ice sheet was both higher and colder than anything in the region today. Its chief nourishment came from maritime air masses that traveled northward and eastward from the Gulf, the Atlantic, and the Pacific. Rising over the high, cold margin of the ice and over wedges of cold air extending out beyond it, these air masses yielded snowfall over a wide belt of glacier surface. The precipitation was thus of two

[17] See discussion in Flint 1943 and Chapter 4 of this book.
[18] See Chapter 19.

types, cyclonic and orographic. At the same time the air masses tended to be deflected generally eastward over the broad southwestern and southern marginal zone of the ice sheet. Additional snowfall in this zone resulted. In consequence the net accumulation in the marginal zone probably exceeded that in the central part of the ice sheet. Hence the southern and southwestern parts of the glacier probably were thicker and higher than the central, northern, and northeastern parts. The resulting differential pressures caused basal ice to flow radially away, from beneath the areas where the ice was thickest, toward the north where the ice was thinner. Thus, although the broad expanse of the ice sheet was kept nearly flat, it was not horizontal. It was highest near its southern and western margins and sloped almost imperceptibly downward toward the north and east (Fig. 50).

The nourishment of the extreme northwestern part of the Laurentide Ice Sheet needs special mention. This region, in northern Alberta and Yukon Territory, is dry today because it lies in the rain shadow of the Rocky Mountains. There are grounds for the belief that during the approach of the expanding ice sheet this region was less dry. Southward shift of the polar front, resulting from reduced glacial-age temperatures, would have brought this country more frequently under the influence of polar air with easterly winds, than it is at present. Although polar easterlies are not moist, they can be expected to have brought more moisture to this territory than it now gets from westerly sources, from which it is barred by high coastal mountains standing across the paths of Pacific air masses. This reasoning helps to explain the nourishment of the northwestern part of the Laurentide glacier.

Probably the growth of glaciers in the high mountains of Greenland, with extreme altitudes of more than 10,000 feet, was contemporaneous with the expansion on Baffin Island or even somewhat earlier. In the interior of Greenland the glaciers merged into piedmonts and finally into an ice cap. On the coasts the valley glaciers and outlet glaciers calved into the sea as they do today. We can picture the thickening glaciers entering Baffin Bay and Davis Strait as forming a floating shelf, similar to the Ross Shelf Ice[19] in the Antarctic, that extended from shore to shore. Because the depths of those water bodies do not much exceed 6000 feet it is even possible that the glacier ice, resting on the bottom, was thick enough to replace floating ice, welding the Greenland Ice Sheet to the glacier ice of the North American mainland. If this condition existed, however, probably it was confined to the phase during which the ice reached its maximum thickness. Whatever the interconnections of this ice, it would have received increments from Hudson

[19] See Chapter 4.

Strait and Labrador and would have discharged bergs into the Labrador Sea, which at that time must have been largely choked with pack ice.

Nourished by moist air masses moving northward, eastward, and northeastward, the growing Laurentide Ice Sheet was most active toward its western and southern margins, where maximum accumulation induced a maximum rate of flow. As it expanded it may have encountered and merged with small plateau ice caps and valley glaciers already formed on highlands in its path. It is likely that such outposts formed in advance of the spreading ice sheet in western Newfoundland, Gaspe Peninsula in southern Quebec, the highlands of New Brunswick, and the White Mountains in New Hampshire. It is possible even that the summer chilling induced by the presence of the growing ice sheet may have been great enough to create perennial snow or even a thin ice cap in the cold continental region northwest of Hudson Bay, where today snow disappears during three summer months. Any such "advance glaciers" would have been overwhelmed by the Laurentide glacier and incorporated in it.

With continued nourishment predominantly over its western and southern parts, the ice sheet expanded westward over the long uphill slope of the Great Plains of Canada nearly to the Rocky Mountains and southward until decreasing latitude brought wastage of the ice into equilibrium with nourishment and rate of flow. The Laurentide Ice Sheet had then reached its maximum extent.[20]

Conditions at Maximum

Thickness of the Ice

The position of the upper surface of the ice sheet has been determined accurately only at places within a few miles of its outer margin, where the upper limit of glaciation can be observed. Farther back from the margin the only data obtainable are minimum figures based on the altitudes of the highest glaciated summits. As most of the glaciated region consists of lowlands and plains the only significant points are the highlands in the east. Some of those known to have been overtopped are shown in the accompanying tabulation.

From these figures can be inferred rough minimum thicknesses for the ice sheet in the vicinities of these highlands, but the figures tell nothing as to the actual thickness. An attempt[21] has been made to

[20] There are both geologic and climatologic grounds for the belief that the ice sheet reached its maximum extent in the southwest before it achieved its full development farther east. This matter is discussed in Chapter 13.

[21] Demorest 1943, p. 391.

Highland	Maximum Altitude (feet)	Height Above Surrounding Country (feet) Maximum	Average
Catskill Mountains, New York	4204	4200	3000
Adirondack Mountains, New York	5344	5200	4000
Mt. Washington, New Hampshire	6288	5000	4500
Mt. Katahdin, Maine	5267	4700	4200
Shickshock Mountains, Gaspe, Quebec	4230	5300	2800
Long Range, Newfoundland	2650	3500	800
Torngat Mountains, Labrador	5000	6000	3000

deduce the thickness from a consideration of the probable altitude of the level of maximum snowfall, which sets an approximate limit to upward growth of an ice sheet. The figure arrived at is 10,000 feet above sealevel. The actual thickness of the ice would be less (or greater) by the value of the altitude of the underlying ground. The thickness must have varied greatly because the western margin of the ice sheet stood at about 5000 feet while its central part, in the Hudson Bay region, rested on a surface that stood well below sealevel. Probably the suggested maximum altitude of 10,000 feet was reached (if it was reached at all) only in the well-nourished southern and southeastern sectors of the ice-sheet margin. Toward the north nourishment was slight, and equilibrium could have been maintained only by the transfer of ice by flow from beneath better-nourished areas. In consequence the ice was thinner in its northern part, but its actual thickness can hardly even be guessed.

A good deal of discussion formerly was provoked by the theory that in the Wisconsin age the Laurentide Ice Sheet was so thin that many highland areas in eastern North America were not covered by it. The basic evidence held to support this belief was chiefly the present-day distribution of plants. Appropriately termed the *nunatak theory*, the concept has been largely disproved.[22]

[22] Prominent among the supposed nunataks were the Torngat and Kaumajet mountains in Labrador, the Long Range in western Newfoundland, the Shickshock Mountains on the Gaspe Peninsula, Quebec, the Keweenaw Peninsula on the south shore of Lake Superior, and the highest peaks in New York and New England. These and other highland areas support isolated colonies of Arctic and Alpine plants that were held to have persisted in these places from pre-Wisconsin time throughout the Wisconsin age, and to have been left isolated there as relicts after deglaciation. Accordingly, it was argued, the Wisconsin Ice Sheet could not have covered those places. Championed mainly by Fernald (1925), this hypothesis found apparent confirmation, according to Coleman (cf. 1921), in the supposed absence of glacially eroded surfaces and glacial deposits in these high places.

In most localities, however, more detailed examination has revealed geologic evidence of glaciation at much higher altitudes than had been recognized formerly. (Cf. Odell 1938; MacClintock and Twenhofel 1940; Flint, Demorest, and Washburn 1942; R. P. Goldthwait 1940). The later work has emphasized the fact that evidence of glaciation on high summits is likely to be small because erosion and deposition by the ice sheet were at a minimum

The Drift Border

As set forth in Chapter 13 the ice did not reach its maximum extent throughout its entire length at the same time. Regardless of the time element each segment of the drift border was controlled in its position by definite factors. On the northeast the glacier ended in the sea. As already suggested, the ice may have been continuous with the Greenland Ice Sheet either as a glacier or as a floating shelf. Off Labrador the ice ended in water, probably as a floating shelf, and was fringed with pack ice.

On the southeast, from the Grand Banks of Newfoundland to New York, the glacier was aground on the continental shelf, ending in an ice cliff and discharging big bergs. Here probably the warm Florida Gulf Stream prevented seaward extension of the ice as a floating shelf. Conditions along this ice front probably were peculiarly foggy and stormy because here the warm Gulf Stream washed the margin of the cold ice sheet. Bergs there were, but probably no pack ice, for the same reason.

On the south and west from the Atlantic to the Rockies the trend of the drift border seems to have been controlled chiefly by topography. Areas topographically low imposed a lobate form upon the ice margin (Fig. 51). From the continental shelf to a point south of the eastern end of Lake Erie the broad Hudson-Ontario lobe formed an irregular arc with its greatest extent around and east of New York City where the land is lowest. West of it the even broader Great Lakes lobe occupied the low territory between the Allegheny Plateau and the Wisconsin highlands. West of the Mississippi the Iowa-Dakota lobe was confined by

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there, and because in such places postglacial frost wedging and other mass-wasting processes occur at a tremendously rapid rate, destroying most of the evidence of glaciation almost as soon as it is exposed by the thinning ice.

Upon detailed examination the botanical evidence, as well as the geologic, fails to support the nunatak theory (Wynne-Edwards 1937). The chief points are these: (*a*) Some of the plants are types demanding a sheltered habitat; they could not have survived on nunataks. (*b*) In some places supposed "relict" plants are shown to have colonized their present positions after the Wisconsin deglaciation. (*c*) A large group of supposed "relicts" are clearly localized by distinct soil and rock types which in some places happen to coincide with highlands. (*d*) There is strong evidence that the whole Arctic-Alpine flora in northeastern North America has been a single unit since pre-Wisconsin time.

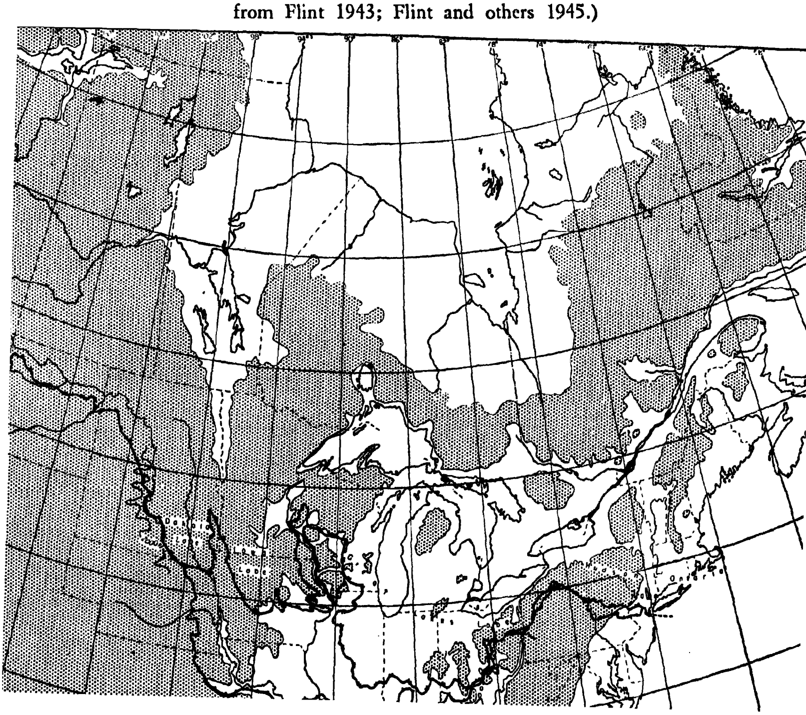
It is concluded that in pre-Wisconsin time the Arctic-Alpine flora of eastern North America occupied much the same places as now, that with the coming of the ice sheet in Wisconsin time it was driven southward, and that during deglaciation it returned northward, ascended highlands, and persisted in those cool places after it had been replaced at lower levels by plants of other types.

The consequence of modern study of the nunatak theory is that the doubt formerly cast on the continuous extent of the Wisconsin Ice Sheet in northeastern North America has been almost entirely removed. That ice sheet now appears to have been nearly if not wholly as thick as its predecessors and only as much less extensive as is indicated by the position of its southern and western drift border.

the shallow north-south topographic sag with an axis along the Red River, hundreds of feet lower than the country to the east and to the west. Farther south this lobe separated into two distinct parts, the Iowa lobe in the Des Moines River and Cedar River basins and a Dakota lobe in the James River basin. The absence of conspicuous lobes on the Great Plains, farther west, is fully consonant with the broad topographic regularity of that region.

From Illinois to the Rocky Mountains the trend of the drift border is strongly north through more than nine degrees of latitude in a distance of less than 1500 miles. Two factors are responsible. First and more important is the rain-shadow effect of the mountains. With increasing proximity to them precipitation decreases, because in this region the precipitation is derived chiefly from air masses crossing the mountains from the direction of the Pacific. The second factor is the steady westward increase in altitude. From the Mississippi to the Rockies the drift border climbs nearly 5000 feet.

FIG. 51. Relation of glaciated area to topography in central and eastern North America. Doubly hatched line = Wisconsin drift border. Singly hatched line = border of the glaciated region. Shaded areas = areas with an altitude of more than 1000 feet. The principal lobes of the Laurentide Ice Sheet at the Wisconsin maximum are named. (Data from Flint 1943; Flint and others 1945.)



The unique re-entrant in the drift border in Illinois and Wisconsin is chiefly a result of topography. It coincides with a kind of peninsula of high land extending northward to Lake Superior and flanked by the much lower Lake Michigan basin on the east and the Iowa lowland on the west. Ice flow was channeled around this re-entrant by the deep basins of Lake Superior and Lake Michigan. The re-entrant was similarly avoided by the pre-Wisconsin ice sheets and is known as the *Driftless Area.* The Driftless Area is surrounded by drift and hence is an island in the drift. But it is surrounded by drifts of various ages and not by any one drift sheet; hence it was never an island in the ice. This fact makes the Driftless Area much less striking than it seems to be when first viewed on a map.

Concerning the western sector of the drift border much is yet to be learned. Through a distance of some 600 miles in Alberta and northeastern British Columbia the ice sheet coalesced with the glaciers flowing eastward from the Rocky Mountains. West of the Mackenzie River in the Northwest Territories the ice was limited by a mountain barrier. Here the glacier "poured westward through the gaps and passes in the eastern flanking ranges of the Rocky Mountains until it reached the barrier formed by the main axial range, when, being unable to pass this, it was deflected to the northwest in a stream from 1500 to 2000 feet deep down the valley of the Mackenzie and thence out to sea."[23] Near the mouth of the Mackenzie the ice sheet, here thin, was confluent with valley glaciers flowing out of the high Brooks Range.

Of the northern limit of the ice sheet very little is known. Plate 3 shows its position as far as existing information permits. It is probable that the glacier was thin, that it ended in a broad floating shelf, and that this in turn was continuous with pack ice in the Arctic Sea — ice more extensive even than it is today. Not until there has been further exploration of the northern fringe of the continent can we be much more definite about the northern terminus of the former ice sheet.

Sea Ice in the Atlantic-Arctic Region

It has already been indicated that in northeastern North America the termini of many glaciers reaching the sea were afloat. If the ice south of the Grand Banks of Newfoundland was afloat at all its extent must have been kept small by the warm Florida Gulf Stream flowing eastward along the coast. Bergs here must have been few and short lived. In the cold water north of Newfoundland the extent of floating shelf ice is wholly conjectural. We lack both direct and indirect evidence as to its

[23] McConnell 1890, p. 543.

seaward extent, though this may have been considerable. It is likely that floating shelves existed along the entire coast of East Greenland.

As for sea ice, there is today continuous pack in wintertime off the entire length of East Greenland and off the Baffin Island-Labrador-Newfoundland coast (Plate 2). During the glacial-age maxima the extent of the pack was certainly much greater. We may expect that the warm North Atlantic Current (the eastward continuation of the Florida Gulf Stream) set an effective southern limit to the pack, as suggested in Plate 3. The pack ice must have been thickly studded with bergs, and some of these, as well as masses of sea ice, may occasionally have floated much farther to the south and east.

Evidence of floating sea ice in the North Atlantic is furnished by the samples of sediments brought up from the deep-sea floor by the Piggot core sampler, which gives a 10-foot core[24] well suited to stratigraphic study. These cores reveal four layers of foraminiferal ooze like that forming on the North Atlantic sea floor today. Interbedded with these are four layers containing sand and pebbles of continental rock types, some of the pebbles having diameters of 2 centimeters and some having faceted surfaces with distinct striations on them. These coarse-grained layers are regarded as having been built up partly by sediment dropped by large numbers of melting icebergs. In contrast the layers of foraminiferal ooze are thought of as representing times when the North Atlantic was at least as free of icebergs as it is at the present day. Alternation of glacial-marine and nonglacial sedimentary layers is indicated further by the fact that the coarse layers contain foraminifera of colder-water types than those present in the fine-grained layers. There is a strong suggestion here of glacial and nonglacial substages or stages, but, because the base of the section has not been reached and the rate of accumulation of the sediments is not known, the correlation of these layers with the glacial drifts on the lands has not been established.

Summary of the Deglaciation

To be able to follow the shrinkage of the Wisconsin Ice Sheet substage by substage around its periphery is only an ideal hope. A discussion so systematic is not possible because of very great gaps in our knowledge. Far more is known about the southern sector of the glaciated region than about all the other sectors combined, and when we try to trace details outward from the well-known sector we quickly run into uncertainty. Accordingly we are obliged to describe the deglaciation by sectors. This is done in Chapter 13 and is illustrated in Fig. 57 and Figs. 52–55. But, in

[24] Bradley and others 1940.

order to keep the sectors related as closely as possible to each other, we shall first present an overall picture of the shrinkage of the ice sheet.

In Chapter 10 it was shown that in central North America four glacial substages within the Wisconsin stage are recognized. The first of these marks the maximum extent of the ice sheet. The second, third, and fourth mark re-expansions of the ice sheet after it had shrunk by an unknown though probably not very large amount. According to the best time estimates the ice sheet uncovered the first or Iowan drift 55,000 years ago, whereas the third or Mankato re-expansion of the ice ended about 25,000 years ago. If these estimates are even approximately correct, deglaciation from the Iowan position to the Mankato position, freeing a belt averaging 300 miles in width and involving repeated re-expansions of the glacier, took a longer time than the subsequent, post-Mankato deglaciation, which freed territory with a radius of more nearly 1000 miles. It follows that deglaciation accelerated throughout its progress. This is an expectable result of the thinning that must have affected the ice sheet, thinning that is recorded by the distribution of eskers as set forth in Chapter 9 and that results in disproportionately large reduction of rate of flow.[25]

Progressive thinning is indicated also by increasing lobation of the ice margin at successive substages.[26] The lobation shows that local topography increasingly controlled the configuration of the ice margin, a circumstance that can only have resulted from thinning.

The evidence of successive drift borders, as far as it goes, suggests that deglaciation was more rapid in the southwest sector of the ice sheet than in the southeast sector. This difference is expectable in that the ice sheet was thinner and less well nourished in its southwestern part than in its southeastern part.

Inward from the drift border of the Mankato substage, no later substages have yet been recognized. The successive positions of the receding ice margin are thus far known only in the Great Lakes-James Bay region, and are determined from the northern limits of successive glacial lakes whose northern shores consisted of the glacier margin itself (Fig. 56).

Both cirques and striations bear testimony that several highland areas, after they had been uncovered by the ice sheet, retained local glaciers.[27] These are the Catskill Mountains, New York (cirque glaciers); Mt. Katahdin, Maine (valley glaciers); the highlands of New Brunswick (residual ice sheet); Shickshock Mountains, Gaspe Peninsula (residual

[25] Demorest 1942, p. 62.

[26] Shown on glacial map (Flint and others 1945).

[27] This concept appears to have been first suggested by Hitchcock. See a discussion by Tarr (1900, p. 445).

ice sheet followed later by cirque glaciers); Long Range, Newfoundland (residual ice sheet followed later by cirque glaciers); coastal mountains of Labrador (cirque glaciers). There are about twelve cirques in the White Mountains, New Hampshire. But there is evidence that these were excavated at some time prior to the incursion of the Laurentide Ice Sheet (they may even have been begun during some pre-Wisconsin glacial age), and that they were not reoccupied by local glaciers after the waning ice sheet had uncovered the mountain summits. The Adirondack Mountains in northeastern New York are marked by a few poorly developed cirques, but there is some doubt whether these were occupied by local glaciers after the mountains had emerged from beneath the ice sheet. Evidence of late north-flowing ice in a part of Quebec not far north of Vermont and New Hampshire[28] suggests a reversal of flow in the main ice sheet, if not a separate residual ice sheet, in the northern New England region.

The apparent anomalies in the distribution of residual highland glaciers in northeastern North America are not likely to find satisfactory explanation until much more detailed information on directions of flow in that region has become available.

Undoubtedly a factor in the isolation of highlands in the region between the St. Lawrence and the New England coast and the development of isolated glaciers on them was the presence of the deep and capacious St. Lawrence trough and its seaward continuation beneath Cabot Strait[29] between Newfoundland and Nova Scotia. The evidence of weak glaciation discussed in Chapter 5, in the highlands immediately south of this trough, coupled with the evidence of strong glaciation by seaward-flowing ice within the trough itself suggest that ice from the basal part of the Laurentide glacier was diverted eastward through the trough. Probably this diversion took place at all times, even during the Wisconsin maximum, because the trough route afforded a smooth outlet with no obstructions. When thinning of the ice sheet began, rapid discharge through this outlet would have led to the rapid appearance of the highlands to the south, as nunataks. In some parts of this region reversal of gradient of the surface of the ice sheet could have resulted, followed by the isolation of ice masses on some of the highlands, independent glaciers thus being formed.

There is some evidence, inconclusive as yet, that, when the ice sheet was at its maximum, centers of radial flow within it existed as far west as northern Alberta. Any such centers not specifically confined to highlands must always have been localized in the outer, best-nourished part

[28] T. H. Clark 1937.

[29] Shepard 1931.

of the ice sheet. They can be expected to have shifted inward as deglaciation progressed. Apparently several such centers existed, during a late phase, immediately west and southwest of Hudson Bay.[30] At this time the ice sheet had shrunk to little more than half its maximum area. It covered the Hudson Bay region and most of Quebec and Labrador as well as Baffin Island. How much territory north of Hudson Bay was still covered by this time is conjectural, though there can be little doubt that much of Ellesmere Island was still under ice.

Shift in effective centers of radial flow farther east is suggested also by striations in Maine. Two sets of striations are evident, an earlier set trending northwest-southeast, and a later set trending north-south. The two sets appear to have been separated in time by an episode of marine deposits in the coastal region. This shows that the margin of the glacier lay well inland when the later set of striations was made. The earlier movement appears to be related to a center in the James Bay region. The later movement seems to have come from central or southern Quebec. The change, though not yet closely dated, probably occurred while the Glacial Great Lakes were developing. A similar shift in the Nova Scotia region is suggested by striations and trends of end moraines.[31]

As deglaciation continued, two distinct centers of radial flow, partly contemporaneous and partly successive, made themselves felt. One was located in the Hudson Bay area; the other lay in the highlands of Quebec, east of James and Hudson bays. Each center is clearly recorded by radial striations and radial eskers.[32] It is not surprising that ice persisted in Quebec because of the high land there, and in the Hudson Bay area because that region represents a southward projection of Arctic climatic conditions and is colder than the region west of it. Hence, after west-central Canada had been deglaciated, resistance of the ice to further wastage would have become stubborn. The ice was continuous across the two areas, though at first the Quebec ice was much thinner than the Hudson Bay ice, both because it rested on a platform 2000 to 3000 feet higher than the floor of Hudson Bay and because it was situated to leeward of the Hudson Bay ice, the presence of which doubtless robbed it of a certain amount of nourishment brought by air masses approaching from the southwest. Ice from the Hudson Bay area flowed southeastward over the area southeast of James Bay.

Slowly, however, the Hudson Bay mass lost substance, and as it did so more nourishment came to the Quebec ice on its high platform. Hence the Quebec ice may actually have increased somewhat, despite the gen-

[30] Data summarized in Flint 1943.
[31] Wickenden 1941.
[32] Flint and others 1945.

eral shrinkage, especially after the surface of the Hudson Bay mass had been lowered to an altitude of less than 2000 feet. The result was a local reversal of the direction of flow in the area southeast of James Bay. Quebec ice flowed southwestward over this area, partly obliterating the striations made here earlier by the southeast-flowing Hudson Bay ice.

It is probable that the ice disappeared from the Hudson Bay region while persisting over the Quebec highlands and the Labrador mountains. It disappeared also from the lowland of western Baffin Island and the Foxe basin, though it still persists on the mountains and plateaus of Baffin, Devon, Ellesmere, and Axel Heiberg islands — all in addition to the persisting ice sheet on Greenland. Presumably deglaciation still has some distance to go before full interglacial conditions will have been attained.

Chapter 13

WISCONSIN STRATIGRAPHY OF NORTH AMERICA

INTRODUCTION

The stratigraphic details of the Wisconsin drift are best known by far in the Central States, where thick clay-rich till and extensive areas of flat land have favored the development of distinct soil profiles and where layers of loess facilitate the tracing of drift sheets. Further, because among the bedrocks easily eroded shales and limestones predominate, the tills are thick and hence have a more distinctive topographic expression than the thin tills in the hilly and mountainous country on the two flanks of the continent. Therefore the stratigraphy of the Central Region is described first, for it sets a standard against which the successions in other parts of North America must be measured.

CENTRAL NORTH AMERICA[1]

In central North America the stratigraphic record of the Wisconsin stage is one of general deglaciation marked by re-expansions of the Laurentide Ice Sheet at irregular intervals. The relative lengths of the intervals are measured by the degree of decomposition of the drifts and by the relation of the tills to the features made by the glacial Great Lakes.

The principal drift sheets — Iowan, Tazewell, Cary, and Mankato — are described first, followed by a summary of the evolution of the Great Lakes. An understanding of both drift sheets and lake features is essential to the picture of deglaciation of the Central Region. The borders of the successive drifts are shown in Fig. 57.

IOWAN SUBSTAGE

The Iowan substage is the earliest Wisconsin drift.[2] Best developed and most extensively studied in Iowa, it has not yet been unequivocally identified outside of Iowa and southern Minnesota. However, there is

[1] The table in Chapter 11 is a useful addition to the discussion of this region.

[2] The Iowan was originally believed to be a distinct glacial stage equivalent in rank to the Illinoian. By some it was thought to be a part of the Illinoian stage. There is now no doubt that the Iowan is an integral part of the Wisconsin stage.

in Wisconsin a drift that is very probably Iowan. To the west, drift that is either Iowan or Illinoian occupies a broad belt of country in North Dakota, eastern Montana, and adjacent parts of Canada.

The Iowan drift is a stony and bouldery clay till. Boulders are larger and more numerous than in the underlying Nebraskan and Kansan tills. Although in western Iowa the Iowan till is very thick, in the eastern part of that State, the region where it has been studied intensively, it is remarkably thin, having an estimated average thickness of less than 10 feet. After having examined this drift in eastern Iowa T. C. Chamberlin once remarked that "it seemed as if a strong east wind might blow it all away." The drift sheet has an "attenuated" border unmarked by end moraine. Both its thinness and its inconspicuous border suggest that in eastern Iowa the ice was very short lived.

In Iowa the Iowan drift represents the maximum southward extent reached by the ice sheet during the Wisconsin age. Farther east, in Illinois and Indiana, the somewhat younger Tazewell drift represents the maximum Wisconsin advance. There the Iowan drift, if it is present, is concealed beneath the Tazewell and other younger deposits. These facts have been interpreted as indicating that two distinct ice sheets existed, one on either side of the Hudson Bay region, and that these ice sheets expanded and shrank alternately.[3] It is more consistent with all the facts to view the Iowan and Tazewell drifts as deposits made by alternate expansions of centers of radial outflow situated near the outer margin of a single Laurentide Ice Sheet, as the amount of snowfall fluctuated from region to region along the ice-sheet margin.[4]

The topography of the Iowan drift sheet as a whole, though largely constructional, in part consisting of long low ridges paralleling the direction of flow of the ice, is not morainic. It is subdued, lacks closed depressions, and in many areas fails to mask the form of the surface on which it lies. The subdued topography results partly from the thinness of the drift and partly from mass-wasting. In addition to the till there is some outwash, and groups of kames occur in places.

Beyond the Iowan drift area is a wide apron or girdle of loess, which in some places overlaps the Iowan itself. The till beneath the loess is chemically almost unaltered, which indicates that loess deposition commenced before or very soon after the Iowan deglaciation began.

[3] Cf. Antevs 1945.
[4] For a discussion of this view see Flint and Dorsey 1945*b*.

Iowan-Tazewell Interval

That some time elapsed between the Iowan and Tazewell sub-ages is proved by the occurrence of a thick sheet of loess between these two drifts. As the Iowan drift beneath the loess is not leached, loess deposition must have begun at least as soon as the Iowan ice had reached its maximum. Where overlain by the succeeding Tazewell drift the loess has a skeleton soil profile, showing that only a short time elapsed between the cessation of effective loess accumulation and the return of the ice. This interval, represented by weathering, was named Peorian[5] after a locality in Illinois. Its very short duration was not fully realized until long after it was named. The loess overlying the Iowan drift in Iowa was originally called Iowan. Subsequently, however, its fossil content was thought to indicate a time of accumulation somewhat later than that of the maximum extent of the Iowan Ice Sheet. Accordingly the loess came to be called Peorian, the name already in use for the zone of decomposition at the top of the loess and immediately beneath the Tazewell drift, in Illinois. This name, however, is awkward, because the loess so called in Iowa represents continuous deposition from Iowan time to Mankato time, since the Tazewell and Cary drift sheets are not known to be present in Iowa. In Illinois, however, this loess can be called Peorian only outside the limits of the Tazewell drift sheet, because at the Tazewell drift border it splits into two layers: a lower, Iowan loess and an upper, Tazewell loess. In Iowa, and in Illinois outside the Tazewell drift these two layers have not been distinguished. Hence the term Peorian, as originally defined, refers to a very short pre-Tazewell interval; but, as applied to loess, it refers to the time between the Iowan maximum and the Mankato maximum. On the whole it would probably be a good thing if the term Peorian were allowed to disappear from the literature, although the name is so convenient in places where the Iowan and Tazewell loesses have not been separated that it is not likely to be abandoned soon.

The fossil content of the Iowan loess, such as it is, yields a consistent picture of climatic conditions at this time. The most abundant fossils are land snails. A study of 56 species from 16 localities in Iowa has made possible sound ecological comparisons with present conditions because all these species are living today. The results, based not on plant fossils but on the snails, suggest that

> the plant geography of the state . . . was little different from that of today except possibly for a greater admixture of [deciduous and (*sic*)] coniferous trees in the forests, resembling the present-day

[5] Leverett 1898*c*.

> forests of Wisconsin and Minnesota. . . . The faunal profiles from the thick loess near the Iowan glacial border show a general change from deep forest conditions during early loess deposition, to forest border or prairie conditions during upper loess deposition. . . . Dense forests persisted back from the glacial border, especially along the main drainage systems. . . .[6]

The evidence of the snails is borne out by fossil trees. In the lower part of the loess in Iowa have been found logs of yew, spruce, and hemlock.[7] At a corresponding horizon near Peoria, Illinois, pollens of spruce, fir, and pine have been found, together with the wood of larch. In the upper part of the loess at that locality are mosses and twigs of larch.[8] These fossils point to cool, moist subarctic forests whose southern limit now stands at least four degrees of latitude north of the localities where the fossil plants occur.

The evidence furnished by fossil mammals is consistent also. The Iowan loess in Iowa has yielded musk-ox, bison, and mammoth, all of which indicate a cool to cold climate. Probably these animals inhabited the open grasslands and forest borders rather than the forested areas themselves.

On the dual basis of this fossil record and the short overall time (probably about 10,000 years), it does not seem likely that the border of the ice sheet receded more than a few hundred miles during the Iowan-Tazewell interval.

Tazewell Substage

The Tazewell drift sheet reaches its greatest apparent extent in northeastern Illinois and central Indiana but continues eastward through Ohio into New York. Probably it is present on the Atlantic Slope, but substages of the Wisconsin drift have not yet been identified there. The Tazewell drift has not yet been recognized west of the Mississippi River.

The drift consists mainly of a stony, clay-rich till. Outwash is abundant, and there are a few small groups of kames. Conspicuous end moraines are present, of which the two most prominent are the Shelbyville moraine, at the margin of the Tazewell drift sheet, and the younger Bloomington moraine. Each of these moraines records a re-expansion of the Lake Michigan glacial lobe following a shrinkage of unknown amount.

A thin sheet of loess, younger than the Iowan loess, covers much of

[6] Kay and Graham 1943, p. 190.
[7] Kay and Graham 1943, p. 171.
[8] Voss 1933.

the Tazewell drift in Illinois and Indiana. Most of this loess appears to be silt reworked from Tazewell outwash. Indeed, in northern Illinois wind-blown sand reworked from outwash contemporaneous with the Bloomington moraine is so abundant that it forms extensive dune areas, some of which completely blanket the outer slope of the moraine.

As indicated previously, the fact that the Iowan drift is most extensive in Iowa whereas the Tazewell drift is most widespread in Illinois and Indiana suggests that the effective center of radial outflow in this sector of the ice sheet shifted eastward between the Iowan sub-age and the Tazewell sub-age.

Tazewell-Cary Interval

The record of the interval between the Tazewell deglaciation and the readvance of the ice margin in Cary time consists of a thin sheet of loess (deposited chiefly during the deglaciation process), a leached zone[9] at the surface of the Tazewell till, and probably also fossil pollen from peat deposits lying on the Tazewell drift. The pollen consistently records, first, an increase of coniferous trees, notably spruce, fir, and pine, to a state of dominance, followed by decline of conifers and the incursion of deciduous forest trees.[10] It is not certain that this pollen record ends with the Cary sub-age because the peat is not overlain by Cary drift. Consequently the times when pollen accumulation began and ended are not fixed.

Cary Substage

The Cary drift reaches its greatest extent in a broad belt extending through Wisconsin, Michigan, and northern Ohio. It continues west into Minnesota, where it disappears beneath the Mankato drift, and east into northern New York. Farther east it has not yet been differentiated from other Wisconsin drifts. The Cary drift includes till (mostly stony) varying according to the bedrocks of the district, much outwash, kame terraces and kames, and a few eskers, especially in Wisconsin and Michigan. The ice-contact features suggest a thinner ice sheet than existed in Iowan or Tazewell time.

The Cary drift includes end moraines in profusion, some of which include much stratified material. Prominent among them are the massive Valparaiso morainic system forming the southern rim of the Lake Michigan basin, and the Mississinawa moraine in northeastern Indiana and Ohio. Many of these moraines are the product of re-

[9] Thornbury 1940.
[10] Voss 1937.

advances of the margin of this sector of the ice sheet. It is therefore clear that a good many fluctuations of the margin occurred during Cary time. The position of the Cary drift reflects a distinct shift of the center of radial flow for this sector of the ice sheet from east to west, as compared with its position in Tazewell time. Also the Cary sub-age witnessed the beginning of the long and complex history of the Great Lakes, preceded by Glacial Lake Wisconsin in the south-central part of that State.

Two Creeks Interval

The interval separating the Cary sub-age from the Mankato sub-age may be appropriately called the Two Creeks interval because the clearest evidence of events during this time is exposed at Two Creeks in northeastern Wisconsin. There the Cary and Mankato drifts are separated by lacustrine silt and clay,[11] peat, and tree stumps in growth position with prostrate logs broken from them and pointing southwest, the whole inclosed in more lacustrine sediments. The trees include spruce, pine, and birch; mosses and several kinds of mollusks are present also.[12] This assemblage suggests a climate like the present climate of northern Minnesota (about three degrees of latitude north of Two Creeks). Evidently the glacier margin retreated from its Cary position to some position well to the north. At Two Creeks a lake was formed and later drained, and trees grew up on its exposed floor. Readvance of the glacier margin in Mankato time re-created the lake ahead of it, submerging the trees. The ice finally overwhelmed the trees, breaking them off and pushing them over into standing water. Then it overrode the lake floor, depositing till above the buried trees.

In the Fox River Valley in southern Wisconsin this horizon occurs at many localities together embracing an area of 500 square miles. At Neenah several acres of standing stumps are exposed, the trees having been snapped off by the advancing glacier. The trees in this region include pine, oak, ash, linden, cedar, and tamarack; moss is present also.[13]

Peat overlying Cary drift in northeastern Illinois contains pollen that records ascendancy of conifers (chiefly spruce and fir) followed by the gradual replacement of conifers by deciduous trees.[14] However, as this peat is not overlain by Mankato drift its date can not be precisely fixed. Although there is reason to believe that its base dates from shortly after

[11] Correlative with Lake Arkona, mentioned elsewhere in this chapter.

[12] L. R. Wilson 1932; 1936.

[13] P. V. Lawson 1902.

[14] Voss 1934.

the Cary maximum, nevertheless the peat may represent not only the Cary-Mankato interval but much of Mankato and later time as well.

In northwestern Minnesota the Mankato till is underlain by a layer of peat and lignite composed chiefly of spruce and tamarack with mosses and fungi.[15] This layer probably represents the Two Creeks interval, though it may possibly be somewhat older.

An interesting detail of this deposit, though not a surprising one, is that many of the pieces of wood show a greatly reduced width of the outermost twenty annual growth rings. The twenty years that preceded the overwhelming of this spruce-tamarack bog by the expanding Mankato ice were cold, with short growing seasons.

Nowhere has the Cary drift been decomposed beneath Two Creeks or Mankato deposits. From this fact it may be inferred that the Two Creeks interval was short.

Mankato Substage

The Mankato drift lies in a continuous belt across the central and northwestern regions, from the Great Plains of Canada through the Dakotas, Iowa, and the Great Lakes region to the Adirondack Mountains in New York, east of which it has not yet been identified. This drift sheet consists of till of variable thickness and with composition changing with the bedrocks, some outwash, ice-contact stratified drift, and extensive deposits of the Glacial Great Lakes. End moraines are numerous and some of them are massive, especially the great Altamont moraine in the Dakotas and Saskatchewan, but end moraine is not present along the entire Mankato drift border. The major end moraines mark conspicuous readvances of the ice-sheet margin. However, not all the end moraines and other glacial topographic features within the area of Mankato drift are the product of Mankato ice. Some of them are Cary features overridden by Mankato ice and thinly veneered with Mankato till.

There is some evidence of lateral overlap of the drifts laid down by adjacent glacier lobes. Overlap is particularly evident in northern Minnesota and Wisconsin, where till derived from the Upper pre-Cambrian rocks of the Lake Superior district to the northeast is red, whereas till derived from the Cretaceous shales and Paleozoic shales and limestones of the plains region to the northwest is gray. These distinct successive overlaps suggest shifts in the amount of net accumulation in two or more centers of outflow in the ice sheet north of the region in which the overlaps occur.

[15] Rosendahl 1943, p. 126.

Soils on the Mankato drift at Minneapolis, Minnesota, are dated through their relation to outwash as having been developed while the ice-sheet margin was still less than 250 miles distant. The plant and animal fossils in these soils indicate an upland forest of white spruce, balsam fir, white pine and birch, with bogs containing tamarack and black spruce, an assemblage found today in extreme northern Minnesota, 200 miles north of Minneapolis.[16]

Deposits made in the Glacial Great Lakes constitute an extensive part of the Mankato drift, for probably the lakes reached their greatest extent during Mankato time. The succession of lakes will now be briefly traced.

From the Wisconsin maximum to the later part of the Wisconsin age the number of species of aquatic mollusks in central North America more than doubled,[17] reflecting a gradually though irregularly ameliorating climate.

THE GREAT LAKES AND OTHER WATER BODIES

Evidence of the Former Lakes

The presence in east-central North America of an extensive system of capacious valleys, dating from preglacial time but enlarged by glacial scour, and depressed regionally under the weight of the ice sheet, led to the development, beginning in Cary time, of the magnificent sequence of glacial lakes whose successors are the modern Great Lakes. The earlier-formed lakes were ponded between high ground on the south and the ice-sheet margin on the north. Later, as the ice melted away, and the land, relieved of the weight of the ice, slowly rose, the lakes fell to lower levels and by gradual stages became confined to the basins they occupy today.

The record of the former lakes is extensive. It lies in wave-cut cliffs and beach ridges that indicate shorelines now abandoned, as well as bars of various kinds and broad lake-floor sediments. At any one former water level these features end abruptly or gradually on the north and east. This northern and eastern limit indicates the approximate position of the margin of the ice sheet at the time when the lake stood at that level. In general the successively lower strandlines extend successively farther north and east, and the lowest ones extend right around the present lakes, showing that when they were made the glacier ice had wasted entirely out of the Great Lakes basins. However, there are exceptions in which strandlines are blanketed beneath lake-floor sedi-

[16] Cooper and Foot 1932.

[17] Baker 1929.

ments. This evidence indicates either temporary re-expansion of the ice sheet, which covered up existing outlets, or upwarping of the crust, which elevated existing outlets. Both these operations would force lake levels to rise and submerge earlier-formed strandlines.

We are now in a position to review the sequence of lakes as synthesized from the work of Gilbert, Spencer, Leverett, Taylor, Goldthwait, Stanley, and others.[18] All these lakes date from Cary and later time, but there is scattered evidence in the Niagara gorge and elsewhere that lakes were formed in this region during the shrinkage of pre-Wisconsin ice sheets also. The details of the earlier lake systems probably can never be deciphered, as the evidence has been largely destroyed.

Sequence of Great Lakes

Lakes Maumee and Chicago

The first of the Wisconsin Lakes, Lake Maumee (Fig. 52), occupied the western part of the Lake Erie basin and drained southwestward through the valley of the Wabash River to the Mississippi. A little later a similar lake, Lake Chicago, occupied the southern end of the Lake Michigan basin. It was held in by the Tinley moraine and discharged southwestward via the Des Plaines River and the Illinois River to the Mississippi. Lake Chicago is represented by three distinct strandlines which show that it fell twice to lower levels. The two episodes of lowering coincided with accelerated discharge through the spillway outlet, accompanied by erosion. They occurred when lakes in the Erie and Huron basins emptied into Lake Chicago. The intervening stable conditions coincided with reduced discharge when the other lakes discharged elsewhere. Lowering came to an end when erosion uncovered bedrock in the floor of the spillway.

Lake Maumee is represented by at least two strandlines. The lower and later level came into existence as the shrinking ice sheet, after several marginal oscillations, uncovered an outlet lower than the Wabash outlet, leading across southern Michigan via Imlay to Lake Chicago. An independent contemporary lake, Lake Saginaw, feeding into Lake Chicago via the valley of Grand River, occupied the district at the head of Saginaw Bay.

Lakes Arkona and Whittlesey

Withdrawal of the ice margin from the peninsula between Lake Huron and Saginaw Bay opened up a wide connection between Lake

[18] The succession is clearly and fully synthesized by Taylor (Leverett and Taylor 1915). The chief modifications since that time have been introduced by Stanley.

Maumee and Lake Saginaw. This caused the level of Lake Maumee to fall, and the two lakes merged to form Lake Arkona (named for the village of Arkona, Ontario), which discharged, as Lake Saginaw had done, westward via the Grand River valley.

At this point the Cary sub-age came to an end. Oscillations of the glacier margin resulted in the conspicuous readvance that built the Port Huron moraine and that marks the beginning of the Mankato sub-age. As this readvance once more covered the peninsula between Lake Huron and Saginaw Bay, the strait that had connected Lake Saginaw with Lake Maumee was destroyed, and the water level in the Erie-Huron basin was raised to form Lake Whittlesey. This lake discharged through a channel at Ubly, Michigan, westward to Lake Saginaw, whose level, controlled by the Grand River outlet, was unaffected by the new development. Lake Whittlesey extended east to the vicinity of Buffalo, where it was held in by the ice. Lake Chicago was still in existence but was increasing in length as the glacier margin retreated toward the north.

Lakes Wayne and Warren

Shrinkage of the ice sheet east of Buffalo at length opened a channel lower than that at Ubly. Accordingly the water in the Lake Erie basin fell to a new low level, that of Lake Wayne, and discharged eastward via the plexus of channels south of Syracuse, New York, to the Mohawk and Hudson rivers. In the Lake Michigan basin, Lake Chicago persisted.

The glacier margin in central New York now readvanced, blocking the Syracuse-Mohawk-Hudson outlet, and the lake water accordingly rose until the strait east of Saginaw Bay was renewed and a single vast lake extended from near Syracuse west to central Michigan, where it discharged through the old Grand River outlet into Lake Chicago. This lake is known as Lake Warren (Fig. 53).

Lake Lundy

Shrinkage of the ice sheet once more opened the Syracuse channels, and the lake water accordingly fell to their level, forming first the ephemeral Grassmere shoreline and quickly constituting Lake Lundy (Fig. 54), whose strandlines also locally carry the names Dana (in New York) and Elkton (in Michigan).

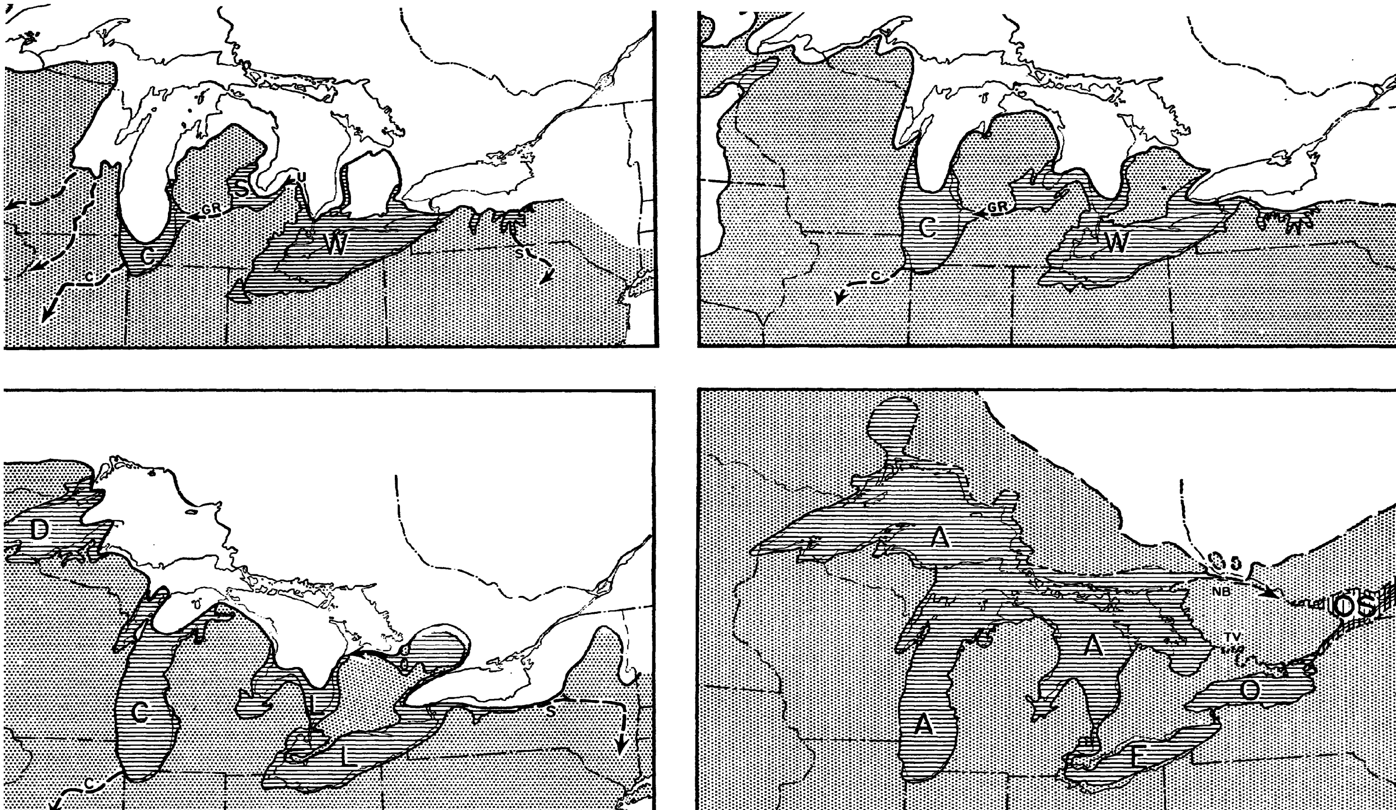
Lake Duluth

While Lake Whittlesey was broadening behind the Port Huron morainic system in the Erie and Huron basins, shrinkage of the glacier in the Lake Superior basin opened a space at the western end of that

TABLE 7. SEQUENCE, MUTUAL RELATIONS, AND OUTLETS OF THE GREAT LAKES
(Outlets are inclosed in parentheses.)

		Superior Basin	Michigan Basin	Huron Basin	Erie Basin	Ontario Basin
Mankato substage	14.	Superior (St. Marys R. to L. Huron)	Michigan (Strait of Mackinac)	Huron (St Clair R to L. Erie)	Erie (Niagara R. to L Ontario)	Ontario (St Lawrence R.)
	13	Nipissing (St Clair R to L Erie, North Bay-Ottawa R to St. Lawrence R.)			Erie (Niagara R to L Ontario)	Ontario (St. Lawrence R.)
	12.	Probable low-water phase			Erie (Niagara R. to L Ontario)	Ontario (St Lawrence R)
	11	Algonquin IV (North Bay-Ottawa R to St Lawrence R)			Erie (Niagara R to L Ontario)	Ontario (St. Lawrence R.)
		(Transition marked by temporary outlets east of Georgian Bay)				
	10	Algonquin III (Trent Valley to L Iroquois, St Clair R. to L Erie, and Des Plaines-Illinois R to Mississippi R , simultaneously)			Erie (Niagara R. to L Iroquois)	Iroquois (Rome channel to Hudson R.)
	9.	Algonquin II (Trent Valley to L Iroquois)			Erie (Niagara R. to L Iroquois)	Iroquois (Rome channel to Hudson R)

Mankato substage	8 Duluth (Huron Mt. channels to L Chicago)	Chicago (Des Plaines-Illinois R. to Mississippi R.)	Algonquin I (St. Clair R. to L Erie)	Erie (Niagara R. to L Iroquois)
	7 Duluth (St. Croix R to Mississippi R)	Chicago (Des Plaines-Illinois R. to Mississippi R.)	Lundy (+ Grassmere, Dana, Elkton) (Syracuse channels to Hudson R)	
	6 Duluth (St Croix R. to Mississippi R.)	Chicago (Des Plaines-Illinois R. to Mississippi R.)	Warren (Grand R. to L. Chicago)	
	5.	Chicago (Des Plaines-Illinois R to Mississippi R.)	Wayne (Syracuse channels to Hudson R)	
	4.	Chicago (Des Plaines-Illinois R to Mississippi R)	Saginaw (Grand River)	Whittlesey (Ubly channel to L Saginaw)
Cary substage	3.	Chicago (Des Plaines-Illinois R to Mississippi R)	Arkona (Grand River)	
	2.	Chicago (Des Plaines-Illinois R to Mississippi R)	Saginaw (Grand River)	Late Maumee (Imlay channel to L Chicago)
	1.			Early Maumee (+ Middle Maumee) (Wabash River)



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FIGS. 52–55. Evolution of the Great Lakes.

FIG. 52. The Great Lakes region at the time of Lake Whittlesey. (Modified after Leverett and Taylor 1915.)

White area = ice sheet.

Ruled areas = glacial lakes (W = Lake Whittlesey; S = Lake Saginaw; C = Lake Chicago).

Arrows = principal lake-overflow stream routes (U = Ubly; GR = Grand River; C = Chicago; S = Susquehanna).

FIG. 53. At the time of Lake Warren. (From Leverett and Taylor, Leverett and Sardeson, W. S. Cooper, and other sources.)

White area = ice sheet.

Ruled areas = glacial lakes (W = Lake Warren; C = Lake Chicago).

Arrows = principal lake-overflow stream routes (GR = Grand River; C = Chicago).

FIG. 54. At the time of Lake Lundy. (From Leverett and Taylor, Leverett and Sardeson, and other sources.)

White area = ice sheet.

Ruled areas = glacial lakes (L = Lundy; C = Chicago; D = Duluth).

Arrows = principal lake-outflow stream routes (S = Syracuse; C = Chicago; SC = St. Croix).

FIG. 55. At the time of Lake Algonquin IV. (From Leverett and Taylor and other sources.)

White area = ice sheet.

Horizontally ruled areas = glacial lakes. (A = Lake Algonquin IV; E = Lake Erie; O = Lake Ontario. Lake Agassiz is not shown.)

Vertically ruled area = area of marine submergence. (OS = Ottawa Sea.)

Arrows = principal lake-overflow stream routes. (TV = Trent Valley (abandoned at this time); NB = North Bay.)

basin in which was impounded Lake Duluth (Fig. 54). This lake spilled southward through the valley of the St. Croix River to the Mississippi. Erosion of this outlet lowered the lake through several levels recognizable in the strandlines. Later the shrinking ice uncovered lower outlets around Huron Mountain near Marquette, Michigan, through which the water, abandoning the St. Croix outlet, drained southeasterly to the now extensive Lake Chicago. Thereafter Lake Duluth fell to the level of Lake Algonquin and merged with it.

Lakes Iroquois, Erie, and Algonquin I

Hitherto the lakes in the Superior, Michigan, and Huron basins had been separate, but now they were approaching a merger into a single Lake Algonquin II, while the Erie and Ontario basins had their own separate water bodies. The change probably began when the shrinking ice uncovered a low channel at Rome, New York. This caused the abandonment of the Syracuse channels by stages, and the lowering of the water in the Lake Ontario basin to the new level, forming Lake Iroquois. This broke the continuity with the water in the Erie basin, which spilled down into the Ontario basin over the Niagara escarpment and thus started Niagara Falls.[19] Separation of the waters in the Erie and Huron basins occurred at the St. Clair River, north of Detroit, and thus an independent Lake Erie came into existence. The water in the Huron basin was thus temporarily isolated as Lake Algonquin I. It discharged south via the St. Clair River into Lake Erie.

Lake Algonquin II

Soon deglaciation of the peninsula separating the Huron and Michigan basins permitted Lake Chicago to fall to the level of, and merge with, Lake Algonquin, thus abandoning the Chicago outlet to the Mississippi. Not long afterward the shrinking ice uncovered a low channel (Trent Valley) through Kirkfield and Fenelon Falls, Ontario. Through this the enlarged Lake Algonquin spilled southeasterly from the Huron basin to Lake Iroquois in the Ontario basin.

Lake Algonquin III

Differential uplift of the crust, caused by removal of the vast weight of the ice sheet, now lifted the northern part of the Lake Alongquin region more rapidly than the southern part. In consequence the lake began to

[19] An earlier Niagara had been instituted in much the same way at one or more times before this latest deglaciation.

rise against its southern shores and once more spilled over both at the St. Clair River and at Chicago. For a short time the Trent Valley outlet competed with these southern outlets, but it soon ran dry. Most of the discharge went via the St. Clair River-Lake Erie-Niagara Falls-Lake Iroquois-Rome-Hudson route. During this time Lake Algonquin reached its maximum area and made the highest shorelines now present in the northern Lake Huron region.

Lake Algonquin IV

Gradual glacier shrinkage successively opened new and lower outlets leading eastward from Lake Algonquin across the highlands east of Georgian Bay to the Lake Ontario basin. In consequence the surface of Lake Algonquin fell successively to lower levels, pausing at each to make a shoreline (Fig. 55). These shorelines are the Wyebridge, Penetang, Cedar Point, and Payette.[20] At length a new outlet, at North Bay, Ontario, was opened leading via the Mattawa and Ottawa River valleys to that of the St. Lawrence, which at that time was submerged to form the Champlain Sea. In the meantime Lake Iroquois had been drained by shrinkage of the glacier to free the northern slope of the Adirondack Mountains, and a brief intermediate stage known as Lake Frontenac had intervened before the creation of Lake Ontario and the gradual encroachment of the sea up the St. Lawrence Lowland.

Nipissing Great Lakes; Possible Low-Water Phase

Continued warping of the crust eventually elevated the North Bay outlet to the level of the St. Clair River outlet, so that for a time the water of the Huron, Michigan, and Superior basins discharged simultaneously through the St. Clair River and the North Bay-Ottawa River channels. This two-outlet phase is termed the Nipissing Great Lakes. At length the North Bay outlet was elevated above the St. Clair River outlet, the former dried up as the Trent Valley outlet had dried up before it, and the Nipissing Great Lakes came to an end. With the entire discharge of the Upper Lakes again passing through the St. Clair River channel, the present Great Lakes system was inaugurated and has continued without major change down to the present time.

Because the Nipissing strandline cuts, essentially with angular unconformity, the more steeply inclined later Algonquin strandlines, and because of the presence of a well-defined submerged stream channel beneath the Straits of Mackinac, it is thought likely that there was a

[20] Stanley 1937, p. 1680.

pronounced hiatus, involving a phase of very low lake levels, in post-Algonquin, pre-Nipissing time.[21] The final solution of this matter has not yet been reached.

The complex changes of level of the ancestral Great Lakes are thus seen to depend on three controls: uncovering and re-covering of outlet channels by the slowly oscillating glacier margin, elevation of outlets by crustal warping, and erosion of the outlet channels by escaping water. Changing combinations of these three factors are responsible for the entire history of the Great Lakes.

Finger Lakes and Champlain Lowland

While the Great Lakes were evolving through their complex history, a similar development on a far smaller scale was taking place on the northern slope of the Allegheny Plateau south of Lake Ontario. The deep north-draining valleys that today contain the Finger Lakes were being uncovered by the ice and filled with meltwater, which in the earlier phases spilled southward over the watershed into the Susquehanna River drainage basin. At first the lakes were independent, each being confined to one valley, but as the ice sheet shrank they coalesced by stages to form a confluent system, held in on the north by the wasting glacier. The highest of these confluent lakes was Lake Newberry, which drained southward over the head of Seneca Lake valley into the Susquehanna basin. This was followed by Lake Hall, which drained westward to Lake Warren in the Erie basin. Next came Lake Vanuxem, which drained eastward past Syracuse to the Hudson, and had two phases caused by shrinkage and re-expansion of the confining ice sheet. Retreat of the ice margin then permitted Lake Lundy, in the Erie basin, to encroach eastward, merging with the Finger Lakes water body (also known at this stage as Lake Dana), and to escape via Syracuse to the Hudson. The opening of Niagara Falls later permitted the water in western New York to fall to the level of Lake Iroquois.[22]

The history of ponded waters in the Champlain lowland east of the Adirondack Mountains began even later in Mankato time than the history of the Finger Lakes. Lake Vermont grew from south to north in the same manner as the Finger Lakes and drained southward into the Hudson Valley, first through a channel at Coveville and later through a lower outlet at Fort Ann. Lake Frontenac, in the Ontario basin, probably was coalescent with a late phase of Lake Vermont. This combined lake was drained as the ice margin shrank away from the south side of

[21] Stanley 1938.
[22] Fairchild 1909, pl. 34–42.

the St. Lawrence Lowland and, as a connection with the sea was opened along the lower St. Lawrence, was succeeded immediately by the marine submergence which, at its maximum, reached as far south as Whitehall, New York.

Marine Submergence of the St. Lawrence Lowland

The evidence of marine submergence of the St. Lawrence Lowland consists of extensive fine-grained deposits containing marine fossils, chiefly invertebrates, and shore features. The recorded history is complex and has not been fully worked out. The basic fact, however, is that owing to the time lag between deglaciation and crustal recovery the St. Lawrence Lowland when deglaciated still lay far below sealevel, so that, as soon as shrinkage of the ice sheet made possible a connection, the lowland filled with water.

There appear to have been at least two distinct marine incursions. The earlier, the Champlain Sea proper, left its mark at the present altitude of nearly 700 feet near Ottawa and possibly flooded a part of the basin of Lake Ontario. As the lagging crustal uplift proceeded, emergence gradually took place to the extent of about 500 feet. Then a long pause in the uplift, probably caused by a climatic change that reduced or halted deglaciation in this region for a time, permitted the rising sealevel, fed by meltwater from wasting glacier ice elsewhere, to record a second incursion. This raised the sealevel in the St. Lawrence Lowland by about 50 feet and constituted the phase that has been called the Ottawa Sea[23] (Fig. 55). This event seems to have occurred during the time of Lake Algonquin IV.

Thereafter renewed upwarping set in, and the St. Lawrence Lowland emerged from the sea to its present position.

LAKES WEST AND NORTH OF THE GREAT LAKES

Lakes Dakota, Souris, and Regina

Early in Mankato time, shrinkage of the Dakota lobe of the Laurentide Ice Sheet from the north-south James River valley in South Dakota opened a lake that lengthened northward after the manner of Lake Chicago and that drained southward over a bedrock threshold near Mitchell. Before it was destroyed by draining, this lake, named Lake Dakota, stretched northward into North Dakota and had a maximum length of 175 miles.[24]

[23] Antevs 1939, p. 716.

[24] Upham 1895*b*, p. 266. Later data suggest this lake was less extensive than originally believed.

At about the same time, in northwestern North Dakota, Lake Souris[25] began to form. This lake was ponded between the massive Altamont moraine on the southwest and the receding glacier margin on the northeast, in the depression now drained by Souris River. Lake Souris first drained southward to the Missouri River, but later, as the shrinking glacier uncovered lower ground, it drained into James River and then into the Red River valley which at that time was occupied by Lake Agassiz. Lake Souris is recorded by many distinct strandlines occupying a vertical span of more than 500 feet. Lowerings of the lake level were caused in part by the uncovering of new outlets and in part by deep erosion of the outlets by escaping water.[26] Although it lay close to the International Boundary, Lake Souris did not occupy Canadian territory until it had dropped close to its lowest and final level.

While Lake Souris was in existence, the basin of the South Saskatchewan River was occupied by a similar glacial lake, Lake Regina.[27] This lake drained through the Souris River into Lake Souris when the latter approached its greatest extent, and later, with continued shrinkage of the glacier and concomitant growth of both Lake Regina and Lake Agassiz, fell to the level of Lake Agassiz and became confluent with it. At its maximum, Lake Regina had a long diameter of 300 miles.

Lake Agassiz[28]

Shortly after the disappearance of Lake Dakota the retreating ice margin uncovered the southern part of the region of the Red River which forms the boundary between Minnesota and North Dakota. Here was impounded a lake, Lake Agassiz I, named after the early advocate of the glacial theory. This water body drained southeastward through the valley of the Minnesota River into the Mississippi at St. Paul. In time Lake Agassiz I grew to so great a size that its area exceeded that of all the existing Great Lakes combined. While the Minnesota River outlet persisted, the lake extended northeastward into Ontario and northward into Manitoba as far as the northern end of the present Lake Winnipeg. When the lake had attained this great extent, it was partially drained (though by what route is not yet known). Then there occurred a conspicuous re-expansion of this sector of the ice sheet that brought the ice margin back almost to the latitude of Winnipeg and renewed the lake (Lake Agassiz II). Evidence of this event consists of (1) the continuity of the highest strandlines, which extend from this latitude through

[25] Upham 1895*b*, p. 267.
[26] Andrews 1939, p. 65.
[27] Johnston and Wickenden 1930. This lake was earlier known as Lake Saskatchewan.
[28] Upham 1895*b*.

250 miles to the outlet via Minnesota River, and (2) a conspicuous unconformity in the lake sediments.[29] The high strandlines of Lake Agassiz thus differ from the high strandlines of the glacial Great Lakes, all of which are short, lengthening northward in the order of decreasing age.

The Minnesota River outlet of Lake Agassiz II persisted until the deglaciation of the southwest slope of the Hudson Bay basin permitted discharge over lower routes. In the difficult country of northwestern Ontario and northeastern Manitoba these routes have not been identified definitely, but there may have been several of them and they may have included the Severn, Hayes, and Nelson rivers, all leading northeastward to Hudson Bay.

Other Lakes in Western Canada

North of Lake Agassiz, other ice-marginal lakes were formed in western Canada, probably at about the same time. Among their modern successors are Lake Athabaska in northern Saskatchewan and Alberta, and Great Slave Lake and Great Bear Lake in the Northwest Territories. Apparently these large water bodies occupy preglacial valleys converted into basins by glaciation. The former lakes are indicated by emerged strandlines and floor deposits, but detailed information on them is not yet available. However, it is probable that when the late Wisconsin history of northwestern Canada has been worked out glacial lakes will prove to have played a large part in it.

Lakes Barlow, Ojibway, and Ojibway-Barlow

While the Nipissing Great Lakes were in existence, continued shrinkage of the ice sheet uncovered the vicinity of Lake Timiskaming, on the boundary between Ontario and Quebec. Here a new lake was ponded between a broad high mass of drift on the south and the glacier margin on the north. This lake, whose shrunken modern remnant is Lake Timiskaming, has been called Lake Barlow.[30] Farther west, in northern Ontario, at about the same time, the waning ice sheet had uncovered the Great Lakes-Hudson Bay divide (the Height of Land). A large lake, Lake Ojibway,[31] was ponded between the glacier and the divide. With continued deglaciation this lake and Lake Barlow enlarged until they merged into a single vast water body which has been called[32] Lake Ojibway-Barlow (Fig. 56). These lakes are recognized, not from strandlines, but from extensive deposits of varved and laminated lake-bottom

[29] Johnston 1916.
[30] M. E. Wilson 1918, p. 143.
[31] Cf. Coleman 1922, p. 40.
[32] Antevs 1925*b*, p. 75.

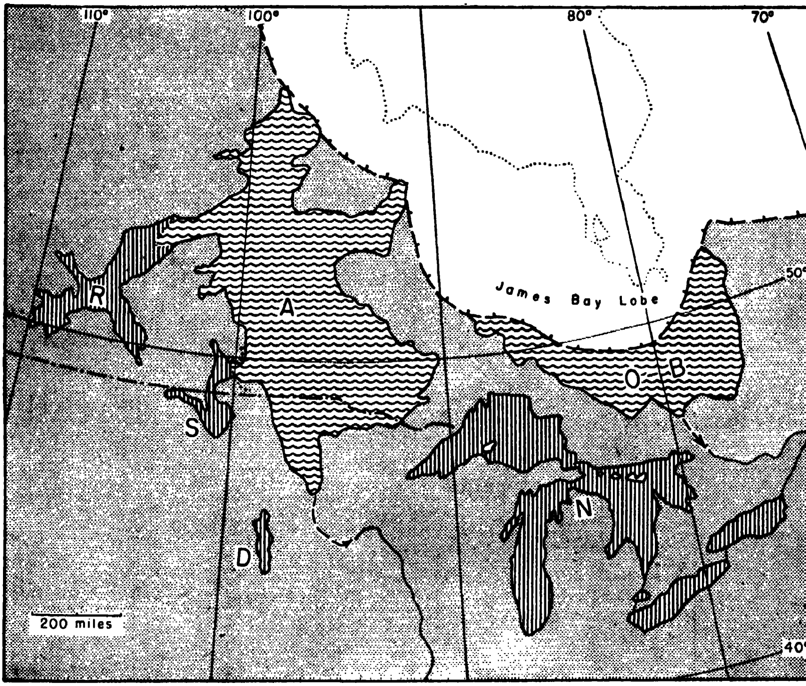


FIG. 56. Northern North America at the close of the existence of Lake Agassiz. (From Geological Survey of Canada and other sources.)

White area = ice sheet.

Horizontally ruled areas = glacial lakes (O-B = Ojibway-Barlow; A = Agassiz).

Vertically ruled areas = glacial lakes of earlier date, drained before this time (D = Dakota; S = Souris; R = Regina; N = Nipissing Great Lakes).

Arrow = southern outlet route of Lake Agassiz.

The areas mapped are the maximum areas of each lake; different levels are not differentiated.

clay and silt, locally containing fossil fishes. The deposits constitute the "Clay Belt" which embraces almost all the tillable land in this part of Canada.

It is possible that, during the later history of Lake Agassiz, that lake merged with Ojibway-Barlow and formed a water body of vast extent, though apparently it was short-lived. The details of these events can not be known until much more information has been gleaned from the difficult country on the Hudson Bay slope of Ontario and western Quebec.

These water bodies were finally drained when ice melting allowed the water to escape northward to Hudson Bay.

Hudson Bay Marine Submergence

Apparently the draining away of the Barlow-Ojibway-Agassiz water body did not occur until the sea had leaked into the Hudson Bay depression, which was then considerably lower than now. Entrance of the sea must have been by way of Hudson Strait. When it first came in, the sea was far more widespread than now because the land stood lower, and formed an extensive blanket of sediments, with many beaches and bars, extending far inland from the present shores of the Bay. It is quite likely that the Barlow-Ojibway-Agassiz water body merged with the invading sea water in the region of what is now James Bay. Continued uplift of the land since the marine submergence was at its maximum has progressively diminished the submerged area to the present area of Hudson Bay-James Bay. Before the uplift has ceased it is probable that almost the whole of this region will have become dry land.

EASTERN NORTH AMERICA

Difficulties of Correlation

In the Appalachians and on the Atlantic Slope the Wisconsin drift has not yet been subdivided. It is fairly certain that the Tazewell, Cary, and Mankato substages are present in western New York, but, although it is known that at several localities farther east two or more Wisconsin drifts occur, that is the sum of our present information. For this state of affairs there are three reasons. First and most important, most of the drift in the eastern region is thin, permeable, rich in quartz, contains insignificant quantities of primary carbonates, and covers terrain having greater relief and steeper slopes than the Central Region. Therefore the methods followed in that region are difficult to apply here. Second, end moraines are much fewer and less conspicuous than farther west. Third, the critical border region from New York City to the Grand Banks of Newfoundland, from which probably the clearest evidence could have been obtained, is submerged beneath the sea.

Despite these serious handicaps it is certain that some subdivision of the Wisconsin stage can be made as a result of detailed study, because scattered evidence has already established the presence of more than one Wisconsin drift. Essential to further study is a successful discrimination between sequences consisting of two tills of distinctly different ages and sequences consisting of a basal till overlain by a superglacial till, both related to a single glaciation. For the present the best we can do is to mention the more promising localities sector by sector and to suggest a

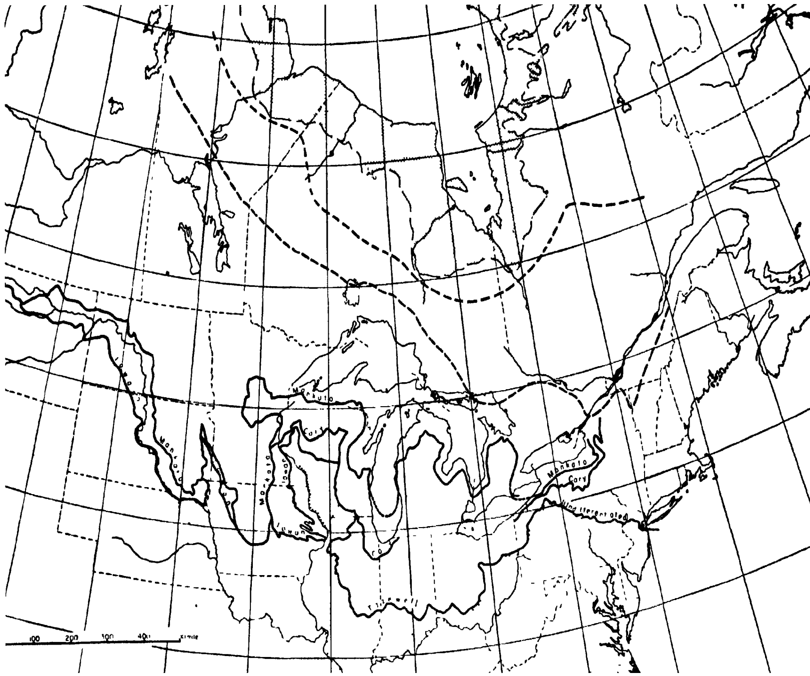


FIG. 57. Successive positions of the margin of the Laurentide Ice Sheet in central North America in the Iowan, Tazewell, Cary, and Mankato sub-ages.

The more southerly of the dashed lines shows the supposed position of the ice-sheet margin at about the time of Lake Algonquin III, the Final phase of Lake Iroquois, and the Coveville phase of Lake Vermont.

The more northerly dashed line shows the supposed position during the closing phases of Lakes Agassiz and Ojibway-Barlow.

Each line represents a roughly but not exactly contemporaneous position of the ice-sheet margin.

For comparison, the dotted line shows the outer limit of the glaciated region; the shaded area is approximately that of the postglacial marine overlap in the Hudson Bay region. (The overlap on the Atlantic coast is too narrow for representation on this scale.) (Compiled from various sources.)

provisional correlation (Table 8), to be corrected and added to as knowledge progresses.

Western and Central New York

In the region in western New York State that centers at Salamanca a sequence of four Wisconsin drifts has been established.[33] The differentiation of these drifts is based not on the presence of zones of decomposition but on their lithologic content, which shows that the ice sheet invaded the region from different directions at different times. The youngest is definitely Mankato, but the correlation of the three older

[33] MacClintock and Apfel 1944.

drifts is uncertain. The correlation suggested in Table 8 is only one of three possibilities.

Farther east, in central New York, a good deal of direct and indirect evidence of plural glaciation has been found, but no correlations have been made.

Northern New Jersey and the Catskill Mountains

Statistical analysis of the degree of decomposition of pebbles of gneiss in the glacial deposits of northern New Jersey suggests that possibly two Wisconsin substages are present, in addition to older drifts that are correlated with the Kansan and Illinoian stages.[34] As the pebbles are more deeply decomposed than those in the drift of northern New England, there can be little doubt that the Wisconsin drift (or drifts) in New Jersey dates from the earlier part of the Wisconsin age.

Farther north, in the Catskill Mountains region, there are two distinct Wisconsin drifts. To the northeast there is a little-eroded drift with conspicuous end moraines, including one at its border. To the southwest is a thicker drift with a more-decomposed boulder content, a subdued surface expression, less-distinct end moraines, and a greater development of postglacial gullies and alluvial fans than in the northern drifts.[35] These facts point to two substages, which, because of their areal position, may prove to be the Tazewell and the Cary.

Long Island and Connecticut

The drift on Long Island is thick, varied, and stratigraphically complex. When it was first thoroughly studied it was believed to include representatives of most of the glacial and interglacial stages recognized in central North America.[36] More recent opinion has tended to the view that fewer stages are present and that the Wisconsin stage is more complicated than was formerly believed.[37] However, aside from a general belief that the Wisconsin drift on Long Island is early, no direct evidence that would serve as a basis for specific correlation with the Central Region has been reported.

North of Long Island, in central Connecticut, old, chemically somewhat altered varved silt and clay are present at several localities and are overlain by much fresher till and a large volume of stratified drift.[38]

34 MacClintock 1940.
35 Rich 1935, p. 120.
36 M. L. Fuller 1914.
37 Fleming 1935; MacClintock and Richards 1936; summary in Flint 1935*a*.
38 Flint 1933.

TABLE 8. SUGGESTED CORRELATION OF PLEISTOCENE

(This correlation is tentative only and should be used in the light of the accompanying
the correlations suggested here. Interglacial

Mississippi Basin		*Western Pennsylvania and New York*	*Connecticut*	*Cape Cod District*
		MacClintock and Apfel 1944	Flint 1933	Woodworth and Wigglesworth 1934; Mather, Goldthwait, and Thiesmeyer 1942
Wisconsin	Mankato	Hamburg and Marilla moraines and associated drift	(Probably not represented)	(Probably not represented)
Wisconsin	Cary	Valley Heads drift		
Wisconsin	Tazewell	Binghamton drift	Hartford lake deposits Latest till	Buzzards Bay and Sandwich moraines
Wisconsin	Iowan	Olean drift	Middletown and Berlin clays; New Haven clay?	Nantucket moraine
Sangamon				*Jacob sand* *Gardiners clay* (*Sankaty beds*)
Illinoian		Illinoian		
Yarmouth				
Kansan		Kansan		Jameco f. (Also Dukes, Weyquosque, and Mannetto members?)
Aftonian				Aquinnah cg.
Nebraskan				
[Pliocene] Citronelle f				

Sequences in Central and Eastern North America

text. The references cited give clear stratigraphic sections, though not necessarily with units are shown in *italic* type.)

Long Island District	*New Jersey*	*Maryland-Virginia-Carolina*	*Bermuda*
M. L. Fuller 1914, MacClintock and Richards 1936 | Fleming 1935	MacClintock and Richards 1936, MacClintock 1940	Flint 1940*b*	Sayles 1931
(Probably not represented)	(Probably not represented)	Low-sealevel cypress horizon near New Bern, N. C.	Eolianite *McGall's soil* Eolianite *Signal Hill soil* Eolianite
Harbor Hill moraine, Ronkonkoma moraine — Latest advance Manhasset f. — Montauk advance, Herod advance	Later Wisconsin Earlier Wisconsin (Budd Lake drift)		
Jacob sand *Gardiners clay*	*Cape May f.*	*Pamlico f.* (*Suffolk scarp*)	*St. Georges soil*
	Illinoian till | Pensauken f. (in part)	Valley trenching	Eolianite
	Pensauken f.	*Surry scarp and related marine deposits*	*Harrington soil* *Marine limestone* *Shore Hills soil* *Marine limestone*
Jameco gravel and contemporaneous (?) old till	* Kansan till | Bridgeton f. (in part) *Bridgeton f.*	Pre-Surry alluvial complex	
Mannetto gravel	Beacon Hill gravel	Brandywine gravel	

From the general position and degree of alteration of the old varved sediments, which are definitely glacial, one might venture the guess that they are Iowan, while the overlying drift is Tazewell. This, however, is no more than a guess until the Wisconsin substages have been traced farther east. There is, further, evidence of re-expansion of glacier ice over the younger drift in Connecticut, but whether it has more than local short-term significance is not known.

MASSACHUSETTS

The Wisconsin stratigraphy of the Cape Cod district is remarkably similar to that of Long Island, and it has been interpreted in much the same way as the Long Island sequence.[39] However, as in Long Island, it seems probable that stratigraphic units originally interpreted as pre-Wisconsin are in fact of Wisconsin age.[40] On the eastern part of Cape Cod a sequence of five tills has been worked out,[41] with a well-marked disconformity separating the upper two from the lower three. The till below the disconformity is oxidized in its upper part. These relations suggest that at least two Wisconsin substages are present. The opinion has been expressed that the uppermost till is "late Wisconsin," a dating that has been attributed also to the surface drift of western Cape Cod.[42]

The relations described or mentioned hitherto are grouped into a tentative correlation chart (Table 8). In the present state of knowledge this chart can not be accurate, but it is reasonable in the light of the facts we have. The Maryland-Virginia-Carolina relations shown on the chart are described in Chapter 17. The Bermuda relations are described in Chapter 16.

In Massachusetts west of Cape Cod the glacial section is less complete and less well exposed, yet two distinct tills, one certainly and the other probably of Wisconsin age, are exposed in many parts of eastern Massachusetts.[43] The older till is deeply oxidized and very compact; the younger till is only superficially oxidized and has a loose texture. In places there are lithologic differences between the two, as though the glacier ice had come from somewhat different directions at different times. Where drumlins are present they are built of the older till, over which is a veneer of the younger till. The younger till is Wisconsin; the

[39] Woodworth and Wigglesworth 1934.
[40] Flint 1935*a*.
[41] Sayles and Knox 1943.
[42] Mather, Goldthwait, and Thiesmeyer 1940, p. 12.
[43] Currier 1941; Chute 1940; Moss 1943. See also F. G. Clapp 1907, p. 514, on tills that may be comparable.

older till is believed to be Wisconsin also. More exact correlations are not yet possible.

Northern New England, Maritime Provinces, and Newfoundland

Information on the far northeastern region is very inadequate because few detailed studies have been made there. At two localities on the southern part of the coast of Maine, sections expose two and three tills,[44] but preliminary study has not succeeded in determining the age of any but the uppermost, which is later Wisconsin. This upper till antedates the marine deposits along the Maine coast and therefore antedates the younger set of striations mentioned in Chapter 12.

Near Joggins in northern Nova Scotia are exposed three tills, differing from one another in their content of rock types but not separated by distinct zones of decomposition.[45] Probably all are Wisconsin, but their exact correlations are as yet unknown.

Near St. Lawrence in southern Newfoundland a fresh till sheet is exposed overlying a decomposed till. Neither deposit has been dated with accuracy.

Postglacial Marine Overlap

From southeastern Massachusetts northward along the east coast and thence westward along the Arctic coast of North America marine deposits overlie the youngest glacial features. These marine sediments, accompanied in many sectors by distinct strandlines marking the positions of former shores, are *prima facie* evidence of postglacial marine submergence of the continental border. As is explained in Chapter 17 submergence was a consequence of the time lag between the deglaciation of any district and the upheaval of the Earth's crust that followed removal of the load caused by the weight of the ice sheet. Coastal districts that had subsided under the loading of the ice sheet to positions below sealevel were inundated as rapidly as the glacier shrank, and after a time they slowly emerged.

While these events were in progress, however, the level of the sea was slowly rising owing to the enormous volume of meltwater poured into it by proglacial streams. Accordingly the position of the strandline in any district at any moment of time was a joint product of crustal upheaval and rise of sealevel. Obviously the marine sediments and shore features at different places date from very different times because of the long time involved in the process of deglaciation. In general the older features occupy the outermost coastal belt; progressively younger features are

[44] Sayles 1927; Sayles and Mather 1937.

[45] Wickenden 1941.

found inland, where embayments such as the St. Lawrence Lowland and the Hudson Bay depression allowed the sea to penetrate far into the interior.

The postglacial marine deposits pass below present sealevel in the latitude of southeastern Massachusetts, along the line at which postglacial rise of sealevel just equals postglacial upheaval of the crust. From this point northward along the coast, marine deposits and related features are discontinuously present. The discontinuities occur in localities where resistant rocks and steep coastal cliffs prevented the formation of marine features or where glacier ice formed the shoreline and thus protected the ground from the sea or where subsequent erosion has destroyed the marine features. On high steep coasts such as Labrador and Greenland the belt of marine deposits is narrow and patchy. On broad areas of low relief such as the Hudson Bay region the belt is wide and more continuous. In some places its width exceeds 150 miles.

The general extent of the postglacial marine overlap in the southern part of the Hudson Bay region is shown in Fig. 57, which is much generalized because except in a few districts information is still scanty. The vertical relations of the marine features in a part of the Eastern Region are indicated in Fig. 81.

In coastal Maine particularly there is evidence that parts of the marine deposits were laid down in the immediate presence of the wasting ice sheet. In many parts of the St. Lawrence Lowland, however, the shores of the long marine estuary[46] that occupied it appear to have been free of glacier ice.

The finer-grained marine deposits quite commonly contain fossils, primarily mollusks of northern types such as live in Arctic and subarctic waters today.[47] Walrus, seals, at least five genera of whales, and various fishes have been found in these marine strata.

GREAT PLAINS REGION

The glacial stratigraphy of the Great Plains Region — the Dakotas, Montana, Saskatchewan, and Alberta — is even less well known than that of the Eastern Region. In the outer (southwestern) part of the glaciated region at least three drift sheets are present. The youngest so far identified is the Mankato drift. Its margin is marked by a massive composite end moraine, the Altamont moraine system, which trends northwest through eastern North Dakota, southern Saskatchewan, and eastern Alberta.

[46] Long known as the "Champlain Sea."

[47] See a brief summary in Richards 1937.

Older drift forms a wide belt southwest of the Altamont moraine, extending to points west of Calgary and Edmonton in Alberta. It includes a large number of discontinuous end moraines, less conspicuous and less fresh in appearance than the Altamont, but the outer margin of it is marked by end moraine only locally. This belt of drift includes at least two members, which are distinct both stratigraphically and topographically. It is probable, though not certain, that both are substages of the Wisconsin drift, though possibly the oldest member is Illinoian. Because so little is known about the glacial geology of this region the known sections, mentioned briefly below, can not yet be correlated.

On the South Saskatchewan River in Saskatchewan and Alberta[48] and its tributary, Red Deer River, in Alberta[49] stratified sediments with peat occur between two tills. The upper till is Wisconsin (earlier than Mankato because of its position southwest of the Altamont moraine), and the lower till may be Wisconsin as well.

In southern Alberta (notably along Oldman River east of Lethbridge) and adjacent northern Montana, beyond the mapped limit of the Mankato drift, are a number of exposures showing what seems to be the same sequence of strata, in some places more than 100 feet thick. This is the generalized section:[50]

Till.
Loess, stratified silt and clay, peat and lignite.
Till, leached and oxidized in upper part.

Both tills consist of material brought from the northeast and are therefore the product of the Laurentide Ice Sheet. The amount of decomposition of the lower till is perhaps slight enough to have been accomplished during one of the intra-Wisconsin intervals. Not only does it seem too slight to be attributable to the long Sangamon age, but the leaching indicates a moister climate than now. Under the present dry climate secondary calcium carbonate is accumulating within 18 inches of the surface. The greater moisture could have resulted only from the presence of an ice sheet to the northeast, and this points to a short intra-Wisconsin interval rather than an interglacial age. This reasoning suggests that the tills are, respectively, Iowan, and Tazewell or Cary, but only detailed studies can decide.[51]

In the Calgary district, which lies farther northwest along the trend of

48 R. T. D. Wickenden, *unpublished.*

49 J. B. Tyrrell 1887, p. 142.

50 Dawson and McConnell 1896; Johnston and Wickenden 1930, pp. 29–44; Alden 1932, pp. 69–77.

51 Johnston and Wickenden (1930, p. 38) suggested that the lower till might be Kansan. This seems very improbable. Alden (1932, p. 71) dated the lower till as not older than Iowan.

the Laurentide drift border, are two Laurentide drifts, both presumably Wisconsin, but differing in topographic expression. One lies to the west and is subdued, with topography that is morainic yet not fresh and sharp. The other lies to the east and has fresh topography with a profusion of conspicuous closed depressions.[52]

At Rolling River, Manitoba (lat. 50° 30′ N; long. 100° 00′)[53] till (probably Mankato) is underlain by a thick section of sand, silt, and clay containing fresh-water mollusks, diatoms, pieces of yew, and seeds of other conifers. As yew grows in bogs in the general region today but also lives under moister and colder conditions, little can be inferred as to the climate that prevailed between the times of till deposition except that probably it was not warmer than now.

In the region south of Regina and Moose Jaw, Saskatchewan[54] Mankato till in or near the Altamont moraine system is exposed at five localities, overlying an older till whose upper part is oxidized and leached. In some exposures the tills are separated by local lake sediments containing fossil water plants and invertebrate animals. The flora and fauna suggest climatic conditions somewhat cooler than now. The interval recorded at these localities may be the Two Creeks interval.

NORTHERN CANADA AND THE CORDILLERAN REGION

The Wisconsin stratigraphy of northern Canada is almost wholly unknown. That vast region has been little explored geologically, and over wide areas the glacial deposits are concealed beneath glacial lake sediments or the postglacial marine overlap. As already indicated, not even the northern limit of glaciation is known.

On the Cordilleran region a good deal of stratigraphic information is available, especially for that part of the region that lies within the United States. However, the glaciation of this region is confined to separate mountain areas, in which local terminology has been used and between which correlation has been attempted only in part. Accordingly, in order to avoid needless repetition, the stratigraphy of the Cordilleran region as a whole, including both Wisconsin and pre-Wisconsin drifts, is dealt with in Chapter 14.

[52] Nichols 1931. These may well correlate with the two tills described above.
[53] J. B. Tyrrell 1892, p. 217.
[54] Wickenden 1931, pp. 65–71.

Chapter 14

PRE-WISCONSIN STRATIGRAPHY OF NORTH AMERICA[1]

The pre-Wisconsin glaciations are best understood by comparison with the Wisconsin glaciation because, although the evidence of them is very incomplete, the facts we have point directly to a close similarity in North America between the Wisconsin glaciers and the pre-Wisconsin glaciers as to origin, character, extent, and related climates. However, we must not fall into the error of assuming that because there are many similarities the course of events was identical during each glacial age. We must be on the alert to discover differences in the evidence that may lead to an understanding of significant differences in the geographic conditions of the successive glaciations.

DRIFT SHEETS OF CENTRAL NORTH AMERICA

Like the Wisconsin drifts, the pre-Wisconsin drifts of America are better developed and much more fully understood in the central part of the continent, the region north of the lower Missouri and Ohio rivers, than in the eastern or western parts, and for the same reasons. The drifts of the Central Region have become, and will probably remain, a standard sequence into which the stratigraphic facts obtained in other regions must be fitted. Accordingly the central region is discussed first.

Possible Pre-Nebraskan Glaciation

The Nebraskan drift is the oldest of which we have any record. It has been generally assumed that because of this fact the Nebraskan glaciation was the earliest or first Pleistocene glaciation of North America. Perhaps it was, but it is at least equally likely that one or more glaciations preceded the Nebraskan, that they failed to reach as far south and west as later drifts, and therefore that their record was buried or destroyed by the later glaciers. There is no direct evidence for or against this concept. But we must be receptive of new evidence that might bear upon it if we hope to bring some light into the present vagueness and obscurity that surround the events connected with the Pliocene-Pleistocene transition.

[1] For general correlation charts see Tables 8 and 9.

Nebraskan Glacial Stage

The Nebraskan drift is best developed in southwestern Iowa[2] and northern Missouri,[3] where it is widely exposed along valleys through erosion of the overlying Kansan drift. In this region the Nebraskan is a massive till with few stones, a matrix rich in clay, and lenses of stratified drift. The large clay content reflects the fact that the regional bedrocks include much shale and other clay-bearing rocks. It also partly explains the very small proportion of outwash deposits which, apparently scanty originally, were still further reduced in bulk by erosion along the stream valleys during the interglacial ages. The thickness of the Nebraskan drift sheet averages 100 feet in Iowa and 50 feet in Missouri, but in places it is believed to reach 200 feet. The Nebraskan deposits cover a dissected bedrock surface with a maximum relief (undoubtedly preglacial) of more than 400 feet. Erosion has removed all traces of whatever morainic topography the Nebraskan till sheet originally possessed. Perhaps because of this erosion, no end moraines of Nebraskan date have been recognized.

This drift occurs also in eastern Nebraska[4] where it is exposed only through erosion of the overlying Kansan drift and other sediments. Here it approaches 100 feet in maximum thickness. Only in northeastern Kansas does the Nebraskan drift extend beyond the limits of later deposits.[5] In this region the till is patchy, and over considerable areas it is represented only by erratic stones and boulders.

Drift believed to be Nebraskan occurs also in northeastern Iowa, eastern Minnesota, western Wisconsin, and southern Ohio. It has been recognized in exposures and borings in western Illinois.[6] At one Minnesota locality the base of the till resting on bedrock contains spruce logs.

Elsewhere this drift has not been identified with certainty, although it is possible that weathered drift and erratics of uncertain date in other parts of the continent may in fact be Nebraskan. It may have been during the time of maximum Nebraskan extent of the ice sheet that the present courses of the upper Ohio River and the Missouri River in the Dakotas were established along the margin of the ice. Their general mode of origin is known, but the date of their origin is obscure.

Only in Missouri, Iowa, Kansas, and Nebraska is the wide extent of the Nebraskan drift clearly discernible beneath the covering of Kansan glacial deposits. In most other peripheral sectors of the glaciated region the Nebraskan ice sheet evidently failed to reach as far south as one or

[2] Kay and Apfel 1929, p. 134.
[3] Holmes 1942, p. 1487.
[4] Lugn 1935, pp. 40–45.
[5] Schoewe 1930, p. 72.
[6] Cf. Wanless 1929.

more of the younger drifts. In consequence the general extent of the Nebraskan ice can only be guessed at. However, comparison (Figs. 58, 59) of the known and inferred southern limits of the Kansan, Illinoian, and Wisconsin ice sheets, respectively, reveals such close general similarity that there can be little doubt that the position of the southern margin of the Nebraskan ice sheet was not far inside the borders of the post-Nebraskan drifts.

Aftonian Interglacial Stage

The Aftonian deposits described originally[7] consisted chiefly of gravel and are not now considered to be evidence of interglacial conditions. However, there have been found features in the same stratigraphic position which clearly indicate a long interglacial interval, and they are referred to as Aftonian. These features consist of gumbotil and layers of peat. Supporting evidence consists of widespread erosion of the Nebraskan drift.

The gumbotil developed on the Nebraskan till sheet, where overlain by the Kansan till, has an average thickness of 8 to 9 feet. Gumbotil of this thickness could have been formed only as a result of long-continued weathering, which, according to estimates based on rate of weathering, lasted about 200,000 years. Where overlain by the Kansan drift, the surface of the Nebraskan drift with its discontinuous capping of gumbotil is an erosion surface marked by conspicuous valleys. These are, however, less deep than the valleys in the preglacial bedrock surface on which the Nebraskan till rests.[8] In some exposures there appears at the top of the Nebraskan till a substance that may be a deeply decomposed loess; in others definite loess lies between the Nebraskan and Kansan drifts.

A picture of the climate during the long Aftonian interval represented by gumbotil and stream dissection is furnished by the pollens contained in five exposures of peat lying between Nebraskan and Kansan drifts in Iowa. The pollens suggest the following sequence:[9]

4. Conifers, bringing the interval to a close.
3. Oak (chiefly in eastern Iowa, and representing a relatively short time).
2. Grasses, indicating a climate much like that of today enduring throughout a long time.
1. Conifers, indicating a long time.

This sequence is regarded as a probably complete record of the Aftonian age.

[7] Bain 1898, pp. 93–98.
[8] Cf. Holmes 1942, pp. 1486, 1487.
[9] Lane 1941, pp. 240, 254–256.

In southern Minnesota much fossil wood imbedded in the Kansan till has been found in deep highway cuts.[10] The wood is well preserved and is preponderantly white and black spruce, though several species of dicotyledons are included. At one locality, sandwiched between Nebraskan gumbotil and Kansan drift, is a layer of peat containing pollens of spruce, balsam fir, pine, hazel (?), maple, and tamarack.[11] Because it overlies gumbotil and because of its northern aspect this layer probably dates from shortly before the arrival of the Kansan ice.

Kansan Glacial Stage

The Kansan drift extends beyond younger drifts in a broad belt through northern Missouri, southern Iowa, northeastern Kansas, eastern Nebraska, northwestern Iowa and adjacent regions, northeastern Iowa, southeastern Minnesota, and Wisconsin northwest of the Driftless Area. Old drift in Pennsylvania and New Jersey and in western Montana have been interpreted as Kansan. Kansan drift underlies the Illinoian drift in southern Illinois and has been identified in an exposure along the Mississippi River near Ste. Genevieve, Missouri. Throughout the region of the Mississippi and Missouri valleys, as might be expected, the composition of this drift is clay till. It closely resembles that of the Nebraskan beneath it, and both reflect the lithology of the regional bedrocks.

As in the Nebraskan drift sheet, lenses of gravel occur in the Kansan till, but neither interbedded gravel bodies nor outwash masses are abundant. This fact is probably attributable to the scarcity of coarse and medium grain sizes in the Kansan till itself, from which the stratified drift must have been derived, and to extensive removal of valley outwash deposits by erosion since Nebraskan and Kansan times.

Throughout the greater part of its region of outcrop, the Kansan till is 40 to 100 feet thick, although in places its thickness exceeds 150 feet. No detailed morainic topography is preserved on the Kansan till sheet; presumably mass-wasting has obliterated it. However, two belts of Kansan end moraine, consisting of broad ridges of thickened drift without morainic topography, occur west and southwest of Lincoln, Nebraska.

The extent of the Kansan drift border is shown in Fig. 58. The most striking feature of its position is its similarity to the borders of the Illinoian and Wisconsin drifts. West of the Mississippi River and in Pennsylvania and New Jersey it was more extensive than the later drifts; in Illinois, Indiana, and Ohio it was less so. In Indiana and Ohio the posi-

[10] Rosendahl 1937.
[11] E. L. Nelson 1935.

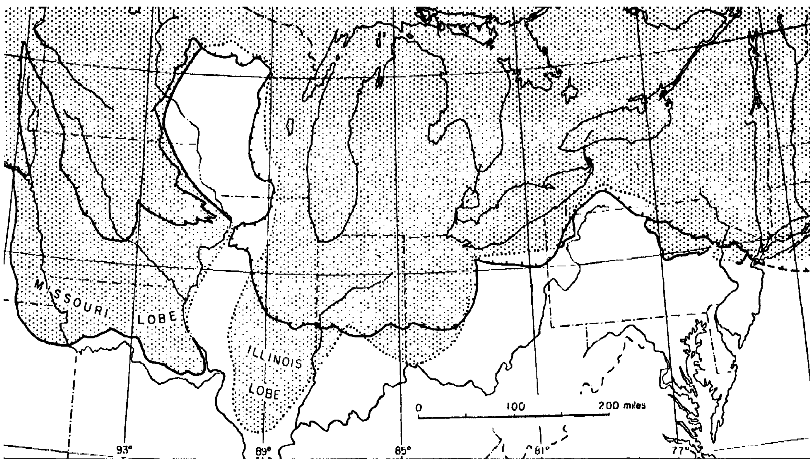


FIG. 58. Border of Kansan drift sheet in central and northeastern United States. (From Flint and others 1945; MacClintock 1933.) Continuous line where drift is exposed; dotted where buried beneath later deposits. Wisconsin drift border (hatched line) is shown for comparison.

tion of its border is unknown, but across Illinois it has been inferred approximately.[12]

The similarity in extent of the Kansan, Illinoian, and Wisconsin drift sheets strongly suggests that in each of the corresponding glacial ages the ice sheets originated in the same districts, developed in the same manner, and were limited in their maximum extent by the same climatic and topographic controls.

Because evidence is scanty, it is not certain that the two prominent Kansan lobes shown on the map (Fig. 58) were not coalescent. If they were indeed separate, as shown, it seems probable that they did not exist, at least in full extent, contemporaneously, or else they would have coalesced. An analogy from the Wisconsin drift sheets is the Iowa lobe of Mankato age, which, extending far southward into Iowa, dates from at least 10,000 years later than the Lake Michigan lobe of Cary age which reached the same latitude. These differences are almost certainly the result of shifting of the loci of maximum precipitation from one sector to another along a marginal belt of the ice sheet.

Nothing is known about the extent of the Kansan drift in the Dakotas, eastern Montana, and the southern plains region of Canada, but this drift should be present in the Plains region.

Yarmouth Interglacial Stage

A long interglacial interval succeeding the Kansan Glacial age is evidenced by the development of gumbotil on the Kansan drift and by peat

[12] MacClintock 1933. Shaw believed that the Kansan drift was thin to begin with and had been largely reworked and incorporated into the Illinoian drift.

beds. The thickness of gumbotil on the Kansan till, where overlain by Illinoian till, averages 11 feet.[13] This figure suggests that the Yarmouth Interglacial age was even longer than the Aftonian age. The length of Yarmouth time has been estimated at 300,000 years. During this time the Kansan drift sheet was dissected, though apparently less deeply than the Nebraskan drift had been dissected during Aftonian time. Many of the present minor streams of northern Missouri appear to have originated on the exposed surface of the Kansan drift. In both Illinois and Iowa some sections reveal loess that is of late Yarmouth age.

The type locality of the Yarmouth stage is Yarmouth, Iowa, where deposits attributed to this stage were identified in the spoil of a dug well. The section was not seen by a geologist, although it was said to include plant remains and bones of skunk and rabbit. However, peat deposits occur between the Kansan and Illinoian drifts at Quincy, Illinois, and at Nilwood, Illinois. Each consists chiefly of pollens of northern coniferous trees (balsam fir, pine, tamarack) and birch, suggesting a climate cooler than now.[14] The restricted range of the vegetation at once suggests that only a small part of Yarmouth time is represented by these sections. A somewhat similar flora in the same stratigraphic position has been reported from Davenport, Iowa.[15] The Yarmouth stage is present as a soil between the Kansan and Illinoian drift sheets at many localities in Illinois.[16]

Illinoian Glacial Stage

As its name implies, the Illinoian drift is best developed in Illinois. From here it occurs almost continuously eastward through Indiana and Ohio and with less continuity through Pennsylvania and New Jersey. Northward and westward this drift extends into eastern Iowa and Wisconsin. Some of the drift older than Mankato that occupies a broad belt in the Dakotas and Montana may be of Illinoian age.

In its type region the Illinoian drift[17] is a till rich in clay though somewhat more stony than the Kansan and Nebraskan tills. Perhaps because of this fact it is accompanied by more gravel than its predecessors, though it is associated with much less gravel than the Wisconsin tills. However, the Illinoian till is enough like the Kansan so that the two can not be distinguished in well logs. Its thickness is extremely variable,

13 Kay and Apfel 1929, p. 241.
14 Voss 1939, pp. 518–523.
15 Baker 1920, p. 248.
16 Leverett 1899, pp. 40–60. The original description is Leverett 1898*b*.
17 A good general reference is Leverett 1899.

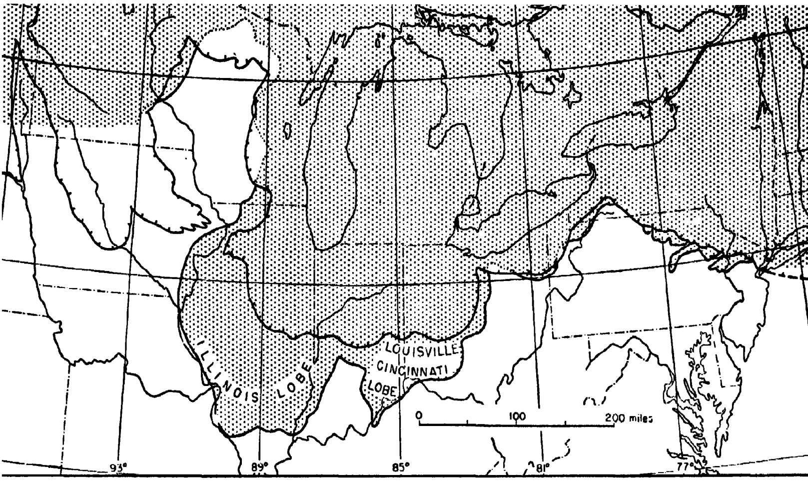


FIG. 59. Border of Illinoian drift sheet in central and northeastern United States. (From Flint and others 1945.) Continuous line where drift is exposed; dotted line where buried beneath later deposits. Wisconsin drift border (hatched line) is shown for comparison.

but on the average it is less than that of the Kansan or Nebraskan till. Toward the north in Wisconsin, and toward the east from Ohio to New Jersey,[18] the Illinoian till becomes much more stony, in consequence of the presence of more resistant bedrocks in these regions.

The Illinoian drift retains morainic topography on broad poorly drained areas, but it is evident that much mass-wasting of swells and deposition in the swales has taken place. Several belts of end moraine, with topographic detail much modified, remain in western Illinois. Stream channels lead away from the Illinoian drift but they do not contain bulky outwash. The reason appears to lie in the very small proportion of coarse rock fragments in the Illinoian till, which consists predominantly of clay and silt. When such material is reworked by meltwater streams of moderate or steep gradient, most of it is exported in suspension. The streams that drained away from the Illinoian ice sheet were probably well loaded with fine sediment which today lies in the Mississippi River delta and beneath the floor of the Gulf of Mexico.

The position of the Illinoian drift border (Fig. 59) is similar in general, though not in detail, to the positions of the Kansan and Wisconsin drift borders. This fact makes it seem likely that the Illinoian ice sheet extended westward over the region of the Great Plains. The ice could hardly have been thick enough to have extended more than 150 miles beyond the position of the Wisconsin drift border in the Mississippi Valley region without having been built far to the west in higher latitudes, unless the climatic conditions were radically different during

[18] MacClintock and Apfel (1944, pp. 1146–1150) have described the Illinoian drift in western New York.

Illinoian time from what they are clearly inferred to have been during Wisconsin time. The Illinoian drift, at present only suspected,[19] in time will probably be found in the Dakotas, Saskatchewan, and adjacent regions.

Sangamon Interglacial Stage

The development on the Illinoian till sheet of a zone of oxidation 10 to 25 feet deep and a zone of gumbotil averaging 4 feet in thickness[20] (where overlain by pre-Iowan loess)[21] records an interval of interglacial weathering that endured, according to a recent estimate, 120,000 years. This interglacial interval was named the Sangamon age because deposits of black muck with pieces of wood, made at that time, are exposed in Sangamon County, Illinois. Interglacial conditions are represented also by erosion of the Illinoian drift sheet, by loess, and by peat.

The best evidence of mild climatic conditions in the Sangamon age comes from a peat bed 20 inches thick exposed at Wapello, Iowa — the only peat bed in Iowa thus far identified as Sangamon. The pollens in this peat consist of northern conifers, except for a zone near the top of the peat in which grasses and oaks are dominant.[22] The grasses and oaks suggest a climate like that at present, but the entire section of peat probably records only a fraction, apparently a late fraction, of Sangamon time.

At Canton, Illinois, a 6-foot layer of peat has yielded a very similar record, with pollens of spruce, fir, and pine, and one zone of oak, beech, and hemlock.[23] As the peat rests on Illinoian gumbotil, it clearly represents only a part of Sangamon time. The peat is overlain by Iowan loess. Exposures of the same horizon containing a comparable flora occur at other Illinois localities.[24]

Near Cattaraugus, in western New York State, a peat believed to be Sangamon contains pollens of spruce, fir, and pine as well as wood and mosses.[25] The northern aspect of this assemblage, together with the fact that the till beneath the peat is decomposed, suggests that the peat dates from just before the earliest Wisconsin glaciation of that district.

Many other deposits, yielding plants and both vertebrate and inverte-

[19] Alden considered it possible that the bulk of the pre-Mankato drift in the Dakotas and Montana is Illinoian.

[20] Kay and Graham 1943, p. 35.

[21] In Sangamon County, Illinois, the age of the loess that overlies the Illinoian drift is in doubt. Consequently the exact upper limit of the Sangamon stage is uncertain and has not been satisfactorily redefined. See Leverett 1898*a*.

[22] Lane 1941, pp. 240, 256–260.

[23] Voss 1939, pp. 523–527.

[24] Voss 1933.

[25] MacClintock and Apfel 1944, p. 1151.

brate animals, are reported by Baker and are provisionally referred by him to the Sangamon age.[26]

The Sangamon deposits also include loess, known in Iowa and Nebraska as the Loveland loess, and in Illinois as Sangamon loess. In Iowa this loess is overlain by the Iowan drift sheet.

Multiple Drifts and Interglacial Horizons of Uncertain Date

Toronto Formation[27]

In addition to the interglacial horizons of known stratigraphic position occurring within the United States, there are several occurrences in cen-

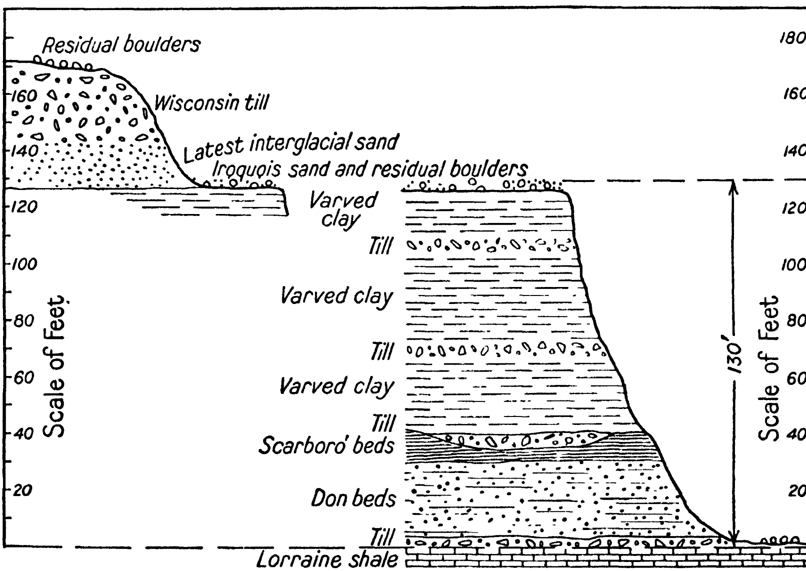


FIG. 60. Sections of the Toronto formation (Coleman 1933). Upper left: section at Leaside brickyard; lower right section at Don Valley brickyard, ½ mile south of Leaside.

tral Canada whose stratigraphic positions are partly or wholly unknown. The best-known of these, and in fact the most remarkable interglacial sequence yet uncovered in North America, is the Toronto formation, exposed at the city of Toronto, Ontario. Known for nearly a century, the deposits have been intermittently studied and have been the subject of many scientific papers, a number of them by A. P. Coleman, whose researches on this interglacial sequence extended over fifty years.

The Toronto deposits are a paradox in that they contain the most complete organic record of warm interglacial conditions so far studied

[26] Baker 1920, pp. 285–343.

[27] A good general reference is Coleman 1933. See also Coleman 1941; Baker 1930.

on this continent; yet their stratigraphic position is as little known as when they were first examined. They present a tantalizing problem to the stratigrapher of the Pleistocene. All told, there are more than a dozen exposures of these beds, scattered through an east-west distance of 20 miles. The best are in the Don Valley brickyard, the Leaside brickyard, and the Scarborough Bluffs of Lake Ontario, east of Toronto proper. The last section, inclusive of related glacial deposits, is more than 350 feet thick. At no single locality is the whole section exposed, but a composite section (Fig. 60) is shown in the accompanying tabulation. (The stratified deposits are referred to lake water that occupied the Ontario basin.)

COMPOSITE PLEISTOCENE SECTION IN THE TORONTO DISTRICT

6. Lake Iroquois (Mankato) features (*wave-cut cliffs, gravel bars, and finer sediments with fossils*).

5 Upper till (*sandy till with swell-and-swale topography, veneering the older deposits*). (*Wisconsin.*)

4. Interglacial sand, possibly marine.

3. Middle complex consisting of tills interbedded with varved fine sediments.

2. Toronto formation (interglacial)
   - Scarborough beds. (*Stratified clay, silt, and sand, with peat. Include 14 species of trees and 72 species of beetles of which 70 are extinct. Both trees and beetles indicate a climate cooler than now, though by no means subarctic.*)
   - — probable conformity —
   - Don beds. (*Lacustrine clay, sand, and gravel, and peat with 32 species of trees, including pawpaw, red cedar, and osage orange; pelecypods and gastropods; vertebrates including woodchuck, deer, bison, bear, and giant beaver, indicating a mean temperature higher than now by 2° to 3° C.*)

1. Lower till, 2 to 3 feet thick; fresh; no weathering of upper surface.

The section is important because of its climatic implications. The flora (represented in part by tree trunks up to 15 feet long) and fauna of the Don beds resemble those of Ohio and Pennsylvania today and thereby indicate a condition of deglaciation at least as extensive as the one that now exists. Possibly even the Greenland Ice Sheet had been greatly reduced or had vanished entirely at the time when these beds were laid down. The Scarborough beds, overlying the Don beds with apparent conformity (though not in the same exposure) indicate a cooler climate and may record the end of interglacial conditions and the oncoming of the succeeding ice sheet. But, beyond the fact that the Toronto formation lies between tills (of which the upper one is Wisconsin) and is therefore definitely interglacial rather than preglacial or postglacial, we have no basis for determining its age. Coleman, who knew the formation better than anyone else, suggested at various times Aftonian, Sangamon, and

Yarmouth dates for it but was never able to find proof of any correlation. It seems likely that the deposits are Sangamon, because they underlie Wisconsin drift and because Illinoian and Sangamon erosion would probably have removed any deposits dating from as early as Yarmouth time.

It is a curious coincidence that the three interglacial deposits better known and climatically more significant than any others on any continent — the Toronto formation, the Eem deposits in the southern Baltic region, and the Hötting deposits in the northern Alps — are stratigraphically alike in being overlain by drifts of the latest glacial age and in resting in turn upon drifts of uncertain age. If the ages of the underlying drifts could be determined, we should be much farther than we are along the road to significant intercontinental correlation of the drifts.

James Bay Interglacial Deposits

The region immediately south and west of James Bay is underlain by weak Paleozoic sedimentary rocks. Probably largely because of this fact, glacial deposits are far thicker there than on the pre-Cambrian rocks farther south and west. In fact Pleistocene deposits are said to be exposed in bluffs along all the rivers in this vicinity. McLearn, the authority on these deposits, says:

> Two drifts at least are present in this area, but only where the interglacial peat deposits are exposed has it so far been possible to distinguish one from the other. Where known the later drift is from 20 to 40 feet thick, whereas the earlier drift is known to be at least 80 feet thick in places.[28]

The interglacial deposits referred to consist of stratified sand and clay, in places including peat, lying beneath till and in some exposures lying between two tills. They occur in the bluffs of the Missinaibi, Opazatika, Soweska, Kwataboahegan, Mattagami, Kenogami, Moose, French, and Abitibi rivers. Those on the first three named, particularly the Soweska, are the most instructive.

The peat layers are thin and consist largely of flattened branches, twigs, and leaves of trees, of which the largest appears to have been no more than 5 inches in diameter, together with *Hypnum* and *Sphagnum* mosses, sedge, willow, and *Equisetum*. Pollen study has shown spruce to be by far the most abundant tree, with pine a poor second, and a scattering of birch and fir. This flora is characteristic of a northern coniferous forest and bog, and records a climate closely similar to that of the same region today.

[28] McLearn 1927, p. 31.

The only indication of a considerable lapse of time is that in one of the exposures along the Soweska River the underlying till is weathered to a depth of 2 feet.

From the data available it is not possible to place these interglacials stratigraphically. It seems likely that they belong either to the Sangamon interglacial stage or, less probably, to an interglacial substage late in Wisconsin time.

Another indication of interglacial conditions in this region is the presence of crumpled clay with marine mollusks underlying till at Moose Factory[29] and along the Albany River.[30] These occurrences suggest that a marine invasion of the Hudson Bay region accompanied a pre-Wisconsin deglaciation, unless (as seems improbable) the clay deposits date from the last deglaciation and were overridden by a minor re-expansion of the remaining ice centered on the highlands of eastern Quebec.

Great Plains Region

Many exposures in the Great Plains region, especially in Canada, show two or more tills separated by fossil-bearing stratified deposits. Probably the majority of these belong to short glacial and interglacial substages within the Wisconsin succession; they are discussed in Chapter 13.

Apparently belonging to a different category is a series of four exposures along Swift Current Creek in southwestern Saskatchewan, at some distance outside the Altamont (Mankato?) moraine. Here the following composite section[31] is exhibited:

	Maximum Thickness (feet)
Till, fresh	70
Sand, gravel, and silt with unidentified plant fragments	23
Till, fresh	7
Sand, silt, and gravel with lignitic beds	72
Till, the uppermost 8 feet, decomposed as to matrix but not as to included stones	83

The 8-foot depth of decomposition of the lowermost till stands in sharp contrast with the very slight degree of weathering found in buried tills elsewhere in southern Saskatchewan, and it suggests, although it does not establish, an interglacial stage rather than merely a substage. The interglacial represented may be the Sangamon. Closely similar sections, probably correlative with the Swift Current Creek sections, occur along the South Saskatchewan River both east and west of Longitude 109°.[32]

[29] Coleman 1941, p. 129.
[30] M. Y. Williams 1921, p. 24.
[31] Wickenden 1930.
[32] R. T. D. Wickenden, *unpublished.*

Also outside the Altamont moraine, near Williston in northwestern North Dakota, is an exposure of two different tills in direct contact with each other.[33] There is no evidence of decomposition of the lower till. The upper till overlies it with erosional unconformity and may have removed a considerable part of the older deposit. Both the texture and the lithologic content of the upper till differ from those of the lower. The exposure lies beyond the Altamont (Mankato?) moraine and beyond a distinctly older pre-Altamont drift sheet, in a district in which the uppermost drift has been doubtfully assigned to the Iowan. If this correlation is correct, it is probable that the lower till at this place is pre-Iowan, perhaps Illinoian. But because of the nature of the evidence the matter is uncertain.

In the Red Deer River region in southern Alberta the orientations of existing and abandoned stream trenches and their relations to each other and to drift fills seem to demand pre-Wisconsin glaciation.[34]

DRIFT SHEETS OF EASTERN NORTH AMERICA

Interglacial Valleys in Central New York

In central New York, evidence of pre-Wisconsin glaciation consists of small interglacial stream valleys. Certain small valleys tributary to the Cayuga Lake valley near Ithaca, broader and deeper than the postglacial valleys in this district, were cut into the side of the main valley after it had been glaciated at least once. Then they were themselves glaciated and filled with till at some time during the Wisconsin age. In widening the Cayuga Lake valley, Wisconsin ice planed away the downstream segments of some of these interglacial valleys and left them hanging conspicuously above the main valley.[35]

A little farther east, at Chittenango Falls State Park, a somewhat similar interglacial valley was identified by Holmes.[36] Here the interglacial valley, excavated in limestone, is 1700 feet long, whereas the postglacial excavation made by the same stream in the same bedrock is 200 feet long. Holmes inferred that the interglacial time represented by the longer valley endured many times longer than the length of postglacial time at this place. This inference points to an interglacial age rather than to a sub-age. No specific interglacial correlation has been assigned either to this valley or to the valleys near Ithaca, but it seems probable that all are referable to the Sangamon age because all the valleys appear to have been glaciated only once since they were formed. Holmes suggested further that a still earlier valley at the Chittenango Falls locality

[33] A. G. Alpha, *unpublished.*
[34] Bretz 1943, pp. 49–52.
[35] von Engeln 1929; Wold 1942.
[36] Holmes 1935.

may date from an early interglacial age rather than from preglacial time.

In western New York the relation of Illinoian till to stream valleys affords indirect evidence of pre-Illinoian glaciation.[37]

Drifts in Pennsylvania

Illinoian Drift

Two pre-Wisconsin drift sheets are recognized in Pennsylvania and New Jersey. In both the upper Allegheny River region in western Pennsylvania, and the Susquehanna River drainage basin in eastern Pennsylvania, drift sheets identified respectively as Kansan and Illinoian[38] extend southward beyond the Wisconsin drift border. In the high Allegheny Plateau country between the two drainage basins no pre-Wisconsin drift has been found. This distribution emphasizes the lobation of the margins of the ice sheets that resulted from the positions of the major relief features encountered by the expanding glaciers.

In the upper Allegheny River region the Illinoian drift is at the surface throughout a belt 85 miles long by about 15 miles wide. In the Susquehanna River drainage basin this drift occurs in a succession of long narrow lobes projecting beyond the Wisconsin drift in the lowlands between the Appalachian ridges. Some of these lobes are as much as 40 miles long by only a few miles wide. In both regions the Illinoian till, where present, is extremely variable in thickness, has a morainic topography which though much subdued is still recognizable, and lacks end moraines. Aside from these points there is little similarity between the two regions. In the Allegheny region the till is coarse-textured and stony, with a very small proportion of fines, and because of its permeability it is oxidized to a rusty brown to depths of at least 30 feet. In the Susquehanna region the till contains much clay and silt. Its content of limestone fragments has been removed by solution, and its contained stones of igneous and metamorphic rock types are deeply decayed. This striking difference in texture is a reflection of differences in the bedrocks underlying the two regions: abundant sandstones in the Allegheny region and abundant limestones and shales in the lowlands of the Susquehanna region. Here as elsewhere the drift takes on the character of the bedrocks, and only a small percentage of its content is recognized as coming from the igneous and metamorphic rocks far to the north.

Illinoian outwash is present in both regions, extending down the Allegheny River to the Ohio and down the Susquehanna to Chesapeake Bay,[39] but it is far more bulky in the former region than in the latter.

[37] MacClintock and Apfel 1944, p. 1161.

[38] Leverett 1934, pp. 94–100, 107–114.

[39] The relation of the Illinoian valley train to the interglacial alluvium near the mouth of the Susquehanna has not been determined, but it is probable that it passes beneath alluvium that is contemporaneous with the (Sangamon?) Pamlico formation.

This difference, too, probably stems from the composition of the bedrocks, and therefore of the tills, in the two regions. As a result of the dearth of clay and silt in the Allegheny region, the till is coarse-textured and so is the bulk of the material that was transported southward by the meltwater streams. The Allegheny River had to transport the cobbles, pebbles, and coarse sand by rolling them along the bottom. In consequence it was obliged to deposit much of this sediment along its upper course in order to build up a gradient steep enough to enable it to carry the remainder out of the region. The sediment that could not be exported constitutes the bulky valley trains whose extensive remnants still fringe the Allegheny River. On the other hand the meltwater streams that fed the Susquehanna carried a far greater proportion of clay and silt, which, moving in suspension, could be exported entirely from the region, leaving the comparatively minor coarse-grained river-bed sediments as the chief constituent of the meager valley trains whose remnants are still preserved. An additional factor that probably contributed to the greater ease of export of outwash by the Susquehanna is that, throughout the extent of the Illinoian valley train, this river had a gradient of 2.5 feet per mile, more than 25 per cent greater than the gradient of the corresponding part of the Allegheny.

Kansan Drift

A pre-Illinoian drift believed to be Kansan extends beyond the Illinoian drift in places in both the Allegheny region and the Susquehanna region. Differences in this drift between one region and the other, although similar to the differences in the Illinoian drift, are not very distinct because the older drift is notably thin and patchy. Over wide areas it consists only of scattered erratic stones and boulders, and exposures of till as much as 10 feet thick are rare. Contained rocks of granitic types are deeply decayed. Outwash is present as fragmentary remnants, and little has been inferred from it.

Drifts in New Jersey[40]

In New Jersey as in Pennsylvania, drifts believed to be Kansan and Illinoian (though together formerly grouped under the name "Jerseyan") form a belt that extends as far as 20 miles beyond the Wisconsin drift border.

The Illinoian drift is very patchy, ranging from spreads of stones and boulders on hilltops to thick deposits of clayey till reflecting in composition the lithology of the local bedrocks. Throughout most of the region the till is no more than 10 to 20 feet thick. It has been considerably modi-

[40] General references: MacClintock 1940; Leverett 1934, pp. 78–90; Salisbury and others 1902.

fied by weathering. Down to an average depth of 12 feet all the pieces of limestone it contains have been removed by solution, and oxidation has proceeded to twice that depth. There is some outwash, though its bulk is not great and it is gravelly only near its source, becoming sandy within short distances away from the drift border. Remnants of it are seen as low terraces along the modern streams.

The Kansan drift is thinner and more patchy than the Illinoian. Scattered stones are more common than till, which is estimated to cover only about 10 per cent of the area of Kansan glaciation beyond the limit of younger drifts. The till is clayey and reaches a maximum thickness of only 30 feet. There is apparently some outwash, but it has not been clearly distinguished from outwash masses of later date that occupy the same group of valleys.

The Kansan drift occurs generally on uplands rather than in valleys; hence it is inferred that the valleys have been cut since the drift was deposited. The Illinoian drift is found in the bottoms of these valleys and is therefore believed to postdate them. Further, pebbles of gneiss in the Kansan drift have been estimated to exhibit, on the average, about 30 per cent more weathering than pebbles of the same gneiss in the Wisconsin drift in the same region.

No end moraines have been identified in either Kansan or Illinoian drift in New Jersey.

Interglacial Alluvial and Marine Deposits[41]

New Jersey

In addition to the till sheets described above, eastern New Jersey has extensive deposits of pre-Wisconsin gravel, sand, silt, and clay that are evidently the remnants of several broad and rather thin bodies of alluvium deposited on the Coastal Plain by streams, notably the Delaware River, flowing southeastward. The distribution of these sediments suggests that each unit was deposited first along pre-existing valleys, that the valleys were gradually filled up, and that the alluvium was then spread over the low intervalley areas of the Coastal Plain until it formed a coalescent, nearly continuous spread whose surface was a broad alluvial plain sloping seaward. The altitudes of the sedimentary units strongly suggest that each was graded to a sealevel higher than the present one, and their mutual stratigraphic relations suggest that, between these times of alluvial filling and lateral spreading, the streams deeply re-excavated the valleys, some of them below present sealevel.

The highest and oldest alluvial unit, the Beacon Hill gravel, exhibits a topographic position and a degree of weathering such that it has always

[41] For further discussion see Chapter 19.

been regarded as preglacial. It is probably the correlative of the Brandywine gravel of Maryland.

The Bridgeton formation, next in age, consists chiefly of gravel and sand. It is thought to have been graded mainly to a high interglacial sea-level. Because it is older than the Pensauken formation, ascribed chiefly to the Yarmouth age, the Bridgeton formation is very likely principally of Aftonian age, though it appears to include some outwash that is probably Kansan.

The Pensauken formation consists of sand and gravel with subordinate silt and clay. Much of it closely resembles the Bridgeton. At one locality it contains a fossil flora that indicates a climate somewhat milder than the climate of the same district at present. Because of this and because of its high altitude the Pensauken formation is believed to have been built in large part by streams that were flowing into an interglacial sea much higher than the present one. Like the Kansan drift in New Jersey, the Pensauken rests in places on hilltops that have been isolated by post-Pensauken valley cutting. Partly for this reason, and partly because it is clearly older than the Cape May formation, the Pensauken formation is thought of as mainly of Yarmouth age, though like the Bridgeton it is believed to include some outwash that may be Illinoian.

The Cape May formation is the youngest of the series and the only member that includes a marine facies. The latter extends up to at least 10 and perhaps 25 feet above present sealevel; the higher and more inland parts of the Cape May formation are alluvial. The fauna of the marine facies records a climate slightly warmer than the present climate along the New Jersey coast. In the vicinity of Staten Island, New York, the Cape May deposits appear to underlie the Wisconsin drift there. Partly for this reason it is generally believed (though not proved) that the Cape May formation dates from the Sangamon Interglacial age.

Thus the Pleistocene stratified deposits in New Jersey appear to record two and perhaps three interglacials. The evidence is far from conclusive, but it is strongly indicative of this interpretation.

Delaware, Maryland, and Northern Virginia

On the Coastal Plain between New Jersey and the James River in Virginia the Pleistocene sequence is similar to the sequence in New Jersey. It is preceded by a somewhat weathered preglacial gravel, the Brandywine gravel, which in position and character, and no doubt in origin and date as well, resembles the Beacon Hill gravel.

The youngest Pleistocene unit thus far recognized in this region is the Pamlico formation. This consists of marine silt, clay, and sand and carries a fossil fauna similar to the Cape May fauna and therefore implying a mild climate. The surface of this deposit is a well-preserved sea-floor

plain rising gently landward from the present shoreline to an altitude of about 25 feet, where it ends abruptly against a wave-cut cliff known as the Suffolk scarp. An alluvial facies extends up the principal streams. The Pamlico formation and the Cape May formation are regarded as contemporaneous in every respect; both are referred to the Sangamon Interglacial.

Between the low-level Pamlico formation and the high-level, preglacial Brandywine gravel there are other deposits of gravel and finer sediments, all of them fluvial and not as yet differentiated from one another. These are probably the equivalents of the Bridgton and Pensauken formations in New Jersey.[42]

Southern Virginia and the Carolinas

The broad marine plain underlain by the Pamlico formation and limited westward by the Suffolk scarp with its toe at 25 feet above sealevel continues southward through Virginia and North Carolina. Inland from it, and extending southward through South Carolina, is a similar plain marked by many spits and bars, underlain by a thin marine formation, and abutting landward against a wave-cut scarp, the Surry scarp, whose toe lies at an altitude of about 100 feet. This scarp is much more extensively eroded than its younger counterpart and is in many places indistinct. Probably the scarp and the deposits to seaward of it are products of the Yarmouth interglacial sea.

The pre-Surry Pleistocene deposits in this part of the Coastal Plain are alluvial, and no record of the high sealevel of Aftonian time has been found in them. It may be that this alluvium met the Aftonian sea along a shoreline that stood somewhat east of the Surry shoreline. In this event all trace of the Aftonian shoreline would have been destroyed during the subsequent encroachment of the sea to the position of the Surry scarp.

Georgia and Florida

The Pamlico formation and the Suffolk strandline continue intermittently around the Florida peninsula on to the Gulf coast. The Surry scarp has not been identified in Georgia and Florida, but the broad territory that extends inland from the crest of the Suffolk scarp up to at least 100 feet above present sealevel is a maze of bars whose component sand constitutes part of a thin surficial layer of marine deposits.

In Georgia and northern Florida marine deposits, with bars including the massive and continuous bar known as Trail Ridge, occur at altitudes higher than those of their upper limits farther north: 160 feet in Georgia

[42] They include sediments called by C. W. Cooke (1935), in his classification, Wicomico, Penholoway, and Talbot, but the validity of this classification has been seriously questioned by Flint (1940*b*; 1941).

and at least 240 feet in Florida. They appear to be inclined northward, passing in eastern Georgia below the altitude of the deposits related to the Surry scarp. The age relation of these high-level features to the Surry features has not been established, but the higher ones are probably the older. They may be deposits made in the high-level Aftonian interglacial sea and afterwards warped into their present position, which they reached prior to Surry (=Yarmouth?) time.

Gulf Coast

Pleistocene marine features are present along the Gulf coast east of the Mississippi River, in some places at altitudes that may exceed 200 feet. Their exact nature and correlation await detailed study. It seems clear that, in the vicinity of the Mississippi River, subsidence resulting from the continuous accumulation of bulky delta sediments has occurred, complicating the shoreline history of that part of the coast. A summary of established facts, together with tentative correlations and a good list of references on this region, has been assembled by Richards.[43]

West of the Mississippi, in Texas, the Pleistocene stratigraphic section consists of the Lissie formation (predominantly sand) and the Beaumont formation (predominantly clay). Each is several hundred feet thick, is dominantly of stream origin, and contains fossils of terrestrial mammals. Each, however, has a marine facies in its seaward part.[44] The fact that both sediments and surface features are mainly of stream origin implies an abundance of stream-borne sediments, the general conditions being analogous to those in New Jersey, already described. However, a faint sea-facing scarp at an altitude of 100 feet is present in places, and there is evidence of quite recent emergence amounting to 20-30 feet.[45] These data suggest that the Surry and Suffolk shorelines may be represented in Texas in sectors in which the volume of alluvium entering the Gulf of Mexico was not too great to prevent some wave cutting.

North of the Coastal Plain the Pleistocene is represented in Texas by terraces and alluvial deposits along the principal streams (Tule, Leona, and Onion Creek formations), in places containing fresh-water mollusks and large vertebrates. More detailed correlations have not yet been made.

Long Island and Cape Cod

There has been much controversy concerning the interpretation of the Pleistocene stratigraphy of this northern, nearly submerged part of the Coastal Plain. Excellent detailed descriptions of Long Island[46] and of

[43] Richards 1939*a*.
[44] Cf. Barton 1930; Richards 1939*b*.
[45] Barton 1930, pp. 1302, 1313.
[46] M. L. Fuller 1914.

the Cape Cod district[47] are available. The present consensus favors an interpretation[48] of the section that differs from earlier views. According to this interpretation there is a preglacial quartzose gravel — the Mannetto gravel on Long Island and its probable equivalent in the Cape Cod district, the Aquinnah conglomerate — that may be the approximate correlative of the Beacon Hill and Brandywine deposits farther south. Above this, both on Long Island and in the Cape Cod district, are two sequences of till and stratified drift separated from each other by the Gardiners clay (Sankaty beds on Nantucket Island). This is a marine deposit, chiefly silt, containing a molluscan fauna similar to that of the Cape May formation. The Gardiners clay is transitional upward into the Jacob sand, which carries a colder-water fauna than the Gardiners fauna, and therefore suggests the oncoming of the succeeding glacial age. The base of the Gardiners clay contains lignitic fossil wood, probably derived from coastal vegetation just before the trees were inundated by the rising interglacial sea.

The sequence of tills and stratified drift deposits ("Manhasset formation") that overlie the Gardiners-Jacob horizon is complex, but there is now wide agreement that the entire sequence is referable to the Wisconsin stage, although it probably represents more than one substage within the Wisconsin. For this reason the Gardiners-Jacob horizon has been referred to the Sangamon Interglacial age, and the underlying complex of tills and stratified drift (Jameco formation) is very probably Illinoian. In the Long Island-Cape Cod region no evidence of a still earlier glaciation has been uncovered.

The Northeastern Region

In eastern North America north of Cape Cod there is little evidence of pre-Wisconsin glaciation. This is not surprising in view of the greater relief, steeper slopes, more highly quartzose lithology, and thinner drift found there as compared with the Central Region. Furthermore, the borders of the successive drift sheets undoubtedly extended from the vicinity of New York City far out upon the continental shelf now covered by shallow water. Hence the present coast between Cape Cod and Newfoundland lies northwest of the drift-border zone in which glacial erosion was at a minimum and in which therefore the chances of preservation of pre-Wisconsin drifts are best.

At a few localities there are exposed two tills, the lower of which suggests a possible pre-Wisconsin age; this is the extent of the evidence. Most of the data on the coast of New England have been assembled by

[47] Woodworth and Wigglesworth 1934.
[48] MacClintock and Richards 1936; Flint 1935b.

Clapp.[49] Other localities are Glendale[50] in western Massachusetts and Rivière du Loup, Quebec.[51]

Greenland

At three places in the Fiord Region of East Greenland, there have been observed valleys whose cross profiles consist of a V valley or narrow U valley incised into a broader V valley, both elements of the profile having been glaciated.[52] The shoulders formed by the junction of the two elements are not the result of differences in resistance of the bedrock, which in all three places is massive gneiss. It seems evident that in each case the narrower element was excavated as a result of rejuvenation of the stream that had created the higher, broader element. It seems improbable that a single glaciation could have so modified preglacial composite valleys as to create forms with very narrow lower members. It is more likely that the lower members were cut out after the upper members had been glacially widened and deepened. If this was so, then there must have been two glaciations separated by an interglacial time when rejuvenated streams excavated the lower elements of these valleys.

This evidence is no more than suggestive, but it should be recorded in view of the unlikelihood of multiple drifts and interglacial deposits ever being found in this region of powerful glacial erosion.

DRIFTS OF THE CORDILLERAN REGION

The stratigraphy of the mountainous region of western North America is difficult to summarize clearly because in most of the mountain districts the evidence of successive glaciations has not yet been thoroughly studied and because rapid erosion in this region of steep slopes has removed a far greater proportion of the evidence of earlier glaciations than has been removed from the lower lands farther east. We have noted that in Iowa the gumbotils, well developed on flat, poorly drained uplands, are developed little or not at all on even the rather gentle valley slopes characteristic of that region. On the much longer and steeper slopes of the Cordilleran mountains mass-wasting and stream erosion are even more effective. In the mountains of central Idaho Capps found pre-Wisconsin till deposits having at the surface a 6-foot zone of soil creep. Within this zone granitic boulders were drawn out into flat shapes several feet in greatest diameter and as little as 1 inch in thickness.[53] Such

49 F. G. Clapp 1907. See also Sayles (1927), who suggests the presence of Illinoian till in southern Maine.

50 Taylor 1910, p. 752.

51 Coleman 1927, p. 391.

52 R. F. Flint, *Glacial Geology and Geomorphology* [of parts of the Fiord Region]. Report of the Boyd Arctic Exped., 1937. Am. Geog. Soc. (in course of publication).

53 Capps 1940, p. 18.

facts as these demonstrate the rapidity of erosion in a highland region.

For these reasons the Pleistocene stratigraphy of the Cordilleran mountains is so little known that except within a few well-studied areas an attempt to separate the drifts would be more misleading than helpful. Therefore both Wisconsin and pre-Wisconsin are here considered together.

Alaska and the Canadian Cordillera

Nearly all the evidence of glaciation in Alaska[54] thus far assembled seems to refer to the Wisconsin stage. In the coastal mountains there are reported to be Pleistocene pre-Wisconsin fluvial, marine, and glacial strata, deformed and lying thousands of feet above sealevel. These relations imply mountain-making movements perhaps contemporaneous with those which affected southwestern California. Near Cook Inlet a deeply decomposed till underlies Wisconsin till, but its age is unknown. The Arctic coastal plain includes marine deposits, perhaps interglacial, up to about 300 feet above sealevel. In places these strata contain a high-latitude molluscan fauna.

The evidence of a long interval of thaw of frozen ground in central Alaska adduced by Taber and mentioned in Chapter 20 surely represents an interglacial interval. So far, however, evidence of only one such long interval has been reported. Capps suggested that in the coastal mountains of Alaska glaciers have existed continuously since the beginning of the Pleistocene epoch; that they expanded during the glacial ages and during the interglacials shrank without disappearing. This is quite possible, though entirely speculative because we do not know as yet the extent of the interglacial rise in temperature in that region. There is room for doubt that temperatures high enough to thaw the ground now frozen in central Alaska and to permit the existence of the ground sloth there would not have destroyed the glaciers in the coastal mountains.

Farther east, in the Carmacks District, Yukon Territory, a gold-bearing alluvium is underlain by a clay-rich till whose included stones (schists and granites) are thoroughly decomposed. In contrast, the stones in the local Wisconsin till are fresh and firm. Probably the older till is pre-Wisconsin.[55]

To the south, in northern British Columbia, evidence of at least two glaciations separated by a long time consists of an episode of deep stream trenching of major valleys, which occurred between two episodes of glaciation.[56] As the interior of British Columbia becomes better known,

[54] General references: Capps 1931; P. S. Smith and Mertie 1930, pp. 236–254; Taber 1943.

[55] Bostock 1936, p. 47.

[56] Johnston 1926, p. 142; Kerr 1934, p. 22.

a much clearer record of its Pleistocene history is likely to appear. This is probable because, as indicated elsewhere, in this region glacial erosion was perhaps less intense than in any other part of the Cordillera, so that, in the valleys and lowlands at least, earlier stages may have been widely preserved beneath a cover of Wisconsin drift.

Puget Sound-Vancouver Region[57]

The Strait of Juan de Fuca and the trough in which Puget Sound lies were occupied by massive lobes of ice that came chiefly from the near-by mountains. The Olympic Mountains between these two lowlands and the Cascade Mountains east of them were the sites of independent glaciers some of which coalesced with the two major lobes.

Three principal stratigraphic members have been recognized thus far. A young till, the *Vashon*, is very fresh, being oxidized to a depth of only 2 feet and containing undecomposed stones. An older till, the *Admiralty*, is oxidized to a depth of 7–10 feet; most of the stones within this depth are decomposed through ⅛ to ¼ inch inward from their surfaces.[58] The older till has a greater extent than the younger, and its vertical distribution shows that it is the product of a thicker glacier. The two tills are separated by stratified sediments (in part marine) and peat. Near Seattle the peat contains an assemblage of plants suggesting a climate somewhat cooler than the present one.[59] The assemblage exposed near Vancouver suggests a climate about like the present one.[60] The stratified sequence contains layers of till and appears to have been deposited during the Admiralty deglaciation. Comparable marine sediments overlie the Vashon till. In the Puget Sound region, the Olympic Peninsula, and on southern Vancouver Island, pre-Admiralty glacial deposits have been recognized.[61] In the Vancouver area the entire Pleistocene sequence has been referred to the Wisconsin stage.

Region of Coalescent Glaciation East of Puget Sound

In the region between Puget Sound and the eastern front of the Rocky Mountains, throughout which glaciers were coalescent, stratigraphic information is scattered and correlations are few. In the Cascade Mountains at latitude 47°30′ three drifts are recognized: (1) a fresh young drift (*Stuart*) confined to high altitudes; (2) a more extensive drift (*Leavenworth*) retaining some morainic topography but oxidized

[57] Good general references are Bretz 1913; C. H. Clapp 1917, pp. 339–355.
[58] J. H. Mackin, *unpublished.*
[59] H. P. Hansen and Mackin 1940.
[60] Johnston 1923, pp. 42–47.
[61] Weaver and others 1944.

through an average depth of 7 feet and with contained granitic stones showing weathering rinds; (3) a still more extensive drift (*Peshastin*) with almost no morainic topography, with oxidation extending through 25 feet, and with the granitic constituents thoroughly decomposed.[62] If the climate of this region be considered not too different from that of Illinois and Iowa to justify rough comparisons of the relative depths of oxidation of the drifts, a correlation of the Leavenworth drift with the Iowan and of the Peshastin drift with the Illinoian may be suggested.

Between the Cascade Mountains and the mountains in northern Idaho no section exhibiting two superposed tills has been reported. However there is evidence, both direct and indirect, of at least two glaciations. The most conspicuous evidence refers to a glaciation that has been called the Spokane and that is very likely an early substage of the Wisconsin. Erosional features and remnants of stream deposits, apparently outwash, bear witness to at least one glaciation of earlier date, and in northern Washington and Idaho there are indications, albeit not conspicuous, of a substage postdating the Spokane.

The Willamette valley in western Oregon contains a deposit of silt in which are imbedded erratic stones believed to have been transported down the Columbia River from sources in eastern Washington during the Spokane episode of great discharge of glacial meltwater. The thickness of weathering rinds in the granitic erratics, and the depth of oxidation of the silt matrix in which they are imbedded, suggest an early Wisconsin (perhaps Iowan) date of deposition.[63]

Principal Separate Glaciated Areas

Information on the separate glaciated areas of western United States (Plate 5) is so widely scattered and for many areas so meager that it seems wise to summarize it in tabular form rather than to attempt a textual summary. Table 6 is a preliminary tabulation of the more important data. Table 9 is a tentative correlation chart suggesting possible time equivalences of the drift sheets identified in the better-studied highlands. From Table 9 it appears that two Wisconsin substages and one (or two) pre-Wisconsin stages have been recognized by various geologists in a number of different highlands. The youngest and least extensive generally recognized drift is fresh both as to topographic expression and as to alteration of the constituents by weathering, and it is commonly thought of as a late-Wisconsin substage. The intermediate drift is more extensive, has a subdued morainic surface, and shows some

[62] Page 1939.
[63] Allison 1935.

chemical decomposition. It is commonly correlated with the Iowan substage. The older drift has a still greater areal extent, possesses little or no morainic topography, is much decomposed, and is separated from its successor by a long interval of uplift and erosion. Opinions as to its correlation differ. Only in the Sierra Nevada has a fourth drift been identified; its age relation to the oldest recognized drift is not known.

Regarding the two Wisconsin drifts two principal possibilities emerge. The first is that more than two glacial ages left their effects in these highlands and that the two drifts recognized are not necessarily the same two drifts in all the highland areas. The recognition of five Wisconsin drifts in the Colorado Rockies might be urged in support of this possibility. The second possibility is that in the mountains of western United States only two glacial expansions of considerable magnitude occurred during the Wisconsin age, and that the other expansions recognized in Central North America are not represented in the west. This general view has been advocated by Antevs.[64] In the present state of knowledge it seems impossible to judge the relative merits of these conflicting ideas. The several Wisconsin drifts reported from the Colorado Rockies do not necessarily carry weight against the second view because it has not been demonstrated that all of them have substage value. Some may represent glacier fluctuations of lesser magnitude.[65]

The pre-Wisconsin drifts involve problems that are even greater. In the Sierra Nevada two pre-Wisconsin drifts have been recognized independently by two authorities in widely separated areas. These authorities are agreed that the older may be Nebraskan and that the younger, represented by massive moraines, may include both the Kansan and the Illinoian stages.[66] Thus far, however, these two stages have not been differentiated in the Sierra Nevada. In the Colorado Rockies and the San Juan Mountains only one pre-Wisconsin drift is recognized, and it is considered to be the same drift in both highlands. Its correlation with the Kansan is very uncertain. In the Uinta Mountains and Yellowstone-Teton-Wind River highlands only one pre-Wisconsin drift has been identified. It has been tentatively correlated with the Kansan stage.

NONGLACIAL DEPOSITS IN WESTERN NORTH AMERICA

Nonglacial deposits and nonglacial stream terraces are present in many parts of nonglaciated North America, but in most places hardly a beginning has yet been made on the study of them. Future thorough studies should make it possible greatly to improve correlations between features

[64] Antevs 1945.
[65] On this point see Jones and Quam 1944.
[66] Matthes 1933, p. 33.

TABLE 9. TENTATIVE CORRELATIONS OF DRIFTS IN SOME

Nonglaciated Regions				
Southern California (*west and south of Sierra Nevada*)	Kansas (*west of the glaciated region*)	Nebraska (*west of the glaciated region*)	Central North America (*for comparison*)	Sierra Nevada West flank (*Yosemite district*)
(H. R. Gale 1931) [There is little agreement on California correlations; *cf* Weaver and others 1944, Barbat and Galloway 1934, Eaton 1941]	(Frye 1945)	(Schultz and Stout 1945)	(Standard)	(Matthes 1930; 1933)
Palos Verdes terrace (*marine, warm water*)	Sanborn formation	Bignell loess "Peorian" loess	Mankato Cary Tazewell Iowan	Late Wisconsin Early Wisconsin
——Older alluvial terraces—— Rancho la Brea beds? (*asphalt with terrestrial vertebrates*)		Loveland formation (*sand, silt, and loess*)	Illinoian	El Portal
—Orogeny, profound erosion— San Pedro zone (*marine, warmer water*) Timms Point zone (*marine, cold water*)	See discussion in reference cited	Upland formation	Kansan	El Portal? (in part)
——Las Posas zone—— (*marine, warm water*) Santa Barbara zone (coast) *Mya Japonica* zone (interior) (*marine, cold water*)		Broadwater formation (*Nebraskan or Aftonian*)	Nebraskan	Glacier Point

in glaciated areas. In the meantime it seems useful only to mention the two regions that have been subjected to intensive study — the Great Plains and southern California.

Extensive though generally thin Pleistocene alluvial and eolian sediments are present in Nebraska and Kansas. Some of them contain assemblages of fossil vertebrates from which some idea of climatic environments can be gained. This region is in a strategic position in regard to correlation because it is crossed by the Platte and Arkansas rivers, both of which originate in the glaciated Rocky Mountains in Colorado and

GLACIATED MOUNTAIN AREAS OF WESTERN UNITED STATES

Glaciated Regions						
Sierra Nevada East flank	San Francisco Mountains	Ruby-East Humboldt Range	San Juan Mountains	Rocky Mountains in Colorado	Uinta Mountains	Yellowstone-Teton-Wind River highlands
(Blackwelder 1931)	(Sharp 1942*c*)	(Sharp 1938)	(Atwood and Mather 1932)	(Ray 1940)	(Bradley 1936) (Atwood 1909)	(Blackwelder 1915 (Horberg 1940)
Tioga	Later Wisconsin	Angel Lake	"Wisconsin"	Sprague Long Draw Corral Creek	Smith Fork	Pinedale
Tahoe	Iowan-Wisconsin	Lamoille	Durango	Home Twin Lakes	Blacks Fork	Bull Lake
Sherwin	Illinoian					
Sherwin? (in part)			Cerro	Prairie Divide	Little Dry	Buffalo
McGee						

thus provide a means of possible correlation with a large area of glaciation on the west. Furthermore the Platte River flows eastward into the region glaciated by the ice sheet, thus providing the possibility of correlation in a second direction. In Nebraska regional correlation has just been begun and is shown in Table 9; in Kansas it has not yet been attempted. It should be possible to carry correlations of such terrestrial deposits across Texas to the coastal-plain sequence already established there.

Southern California has a thick sequence of marine sediments containing fossil invertebrates that show alternation of cold- and warm-

water conditions. General agreement as to the correlation of the several units has not been reached by California geologists. Table 9 shows a correlation that is accompanied, in the original reference, by a detailed and thoughtful discussion of the evidence. Agreement is general on one point, that in this region Pleistocene sedimentation was interrupted by orogeny that folded earlier Pleistocene strata and lifted them thousands of feet above sealevel. Marine terraces postdate the orogeny, but since they themselves are warped their present positions relative to sealevel have little or no value for inferences as to sealevel fluctuations.

At the western border of the city of Los Angeles occur deposits which though confined to an area of less than 1 square mile nevertheless have an important ecological significance. These are the Rancho la Brea beds, a mass of sand, silt, and clay thoroughly impregnated with tar that seeped upward from underlying oil-bearing strata and hardened as a result of slow evaporation. These deposits are filled with the bones of fossil mammals and birds and the remains of coniferous trees. The place is the site of a formerly active oil seep, to which animals came, perhaps thinking it a pool of water, and were entangled in the sticky oil. Carnivorous animals fed upon the trapped carcasses and in turn were caught. The bones of hundreds of individuals have been collected, and many thousands more are believed to be still in the tar awaiting excavation.

The rich fauna, including 39 species of mammals and 60 kinds of birds, has always been regarded as interglacial, though in the absence of any means of close correlation the specific interglacial has not been determined. The recent trend of opinion places the Rancho le Brea beds in the Sangamon Interglacial.[67] Similar seeps at McKittrick and at Carpinteria, California, yield somewhat similar faunas that have been regarded as slightly younger, probably Wisconsin.[68]

Other Pleistocene nonglacial deposits occur in many parts of western United States. Some of them are lacustrine and bear witness to conspicuous climatic fluctuations; they are discussed in Chapter 20. Many others remain to be fully studied and fitted into the stratigraphic sequence of the Pleistocene.

[67] H. R. Gale 1931, p. 73; see also Schultz 1938*b*, p. 156.
[68] Schultz 1938*b*.

Chapter 15

THE FORMER GLACIERS OF EUROPE

INTRODUCTION

During the Fourth Glacial age events in Europe were similar in many ways to the events that took place in North America during the Wisconsin age. In northern Europe, between latitudes 50 and 70, maritime air masses built up an ice sheet that got its start in the Scandinavian Mountains and gradually spread over an area of 1,650,000 square miles. Elsewhere glaciers developed in highlands wherever the combination of altitude and atmospheric moisture was adequate. Chief among these highland areas was the mass of the Alps, which, at the height of glaciation, developed a glacier complex that approached the condition of a mountain ice sheet. Glaciers formed likewise in high parts of the Pyrenees, the Carpathians, the Apennines, and on scattered highland areas in the Balkans, central Europe, France, the Iberian Peninsula, and Britain.

The records of earlier glacial ages show that at those times the conditions were much the same. Glaciers formed in the same highland areas and flowed outward along much the same paths. The chief difference lay in the extent of the earlier glaciers, which in general was greater than that of their latest successors. At its maximum, during the glacial age that preceded the latest one, the Scandinavian Ice Sheet had an area of about 2,145,000 square miles—30 per cent greater than that of its successor.

SCANDINAVIAN ICE SHEET

The Scandinavian Mountains

The Scandinavian Mountains are the backbone of the Norwegian-Swedish peninsula, reaching nearly from end to end of its 1000-mile length. Their crest commonly reaches altitudes of 4000 to 5000 feet, and in southern Norway the highest peaks reach 8500 feet. The mountains are carved from a broad erosion surface upheaved at or shortly before the beginning of Pleistocene time. They form a narrow chain, lying oblique to the paths of warm, moist maritime air masses that approach the Norwegian coast from the southwest. These air masses precipitate

copious moisture on the coastal flank of the mountains at a mean annual rate increasing from 30 inches at the north end to more than 120 inches in places near the southern end. Precipitation is roughly proportional to the height of the mountains. Much of it is rain, but there is enough snowfall, coupled with sufficiently cool and cloudy summers, to maintain plateau ice caps and valley glaciers with a combined present area of 2416 square miles.

Growth of the Ice Sheet

The Scandinavian Ice Sheet had its origin in these mountains, not only in the Fourth Glacial age but in earlier glacial ages as well. The mountains harbor glaciers at present, whereas the adjacent lower regions do not. Drift derived from the rocks of the mountains occurs east, south, and west of the mountain axis.

A comparatively slight reduction of the mean annual temperatures now prevailing in this region would result in conspicuous growth of the glaciers into a semi-coalescent glacier complex, in which plateau ice caps, occupying broad remnants of the upheaved erosion surface, would be conspicuous. Probably this is what happened at the beginning of the Fourth Glacial age. The reduced temperature, lowered regional snowline, and increased expanse of ice and snow with its great heat-reflective power conspired to extend the glaciers[1] and to produce further coalescence.

Snow accumulated in greatest quantity on the windward (western) flank of the mountains just as it does today, although much snow, augmented perhaps by wind-drifted snow, accumulated also on the leeward (eastern) flank. This difference resulted at first in a greater development of glacier ice on the western side of the mountains. However, the removal of ice from that side by glacier flow was far greater than removal from the eastern, much as in the coastal mountains of British Columbia, and for the same reasons. On the steep west slope the ice reached deep water within a comparatively short distance. There over wide sectors its margin was afloat, and calving prevented any great seaward extension of the floating ice. In contrast the ice that flowed eastward followed a longer and gentler slope and, except perhaps locally, was not subject to calving along its terminal margin. Also, extension eastward may have been favored somewhat by drifting of snow by the wind from west to east. These relations are suggested in Fig. 61.

Although snow accumulation probably was greater on the western side of the Scandinavian Mountains than on their eastern side, the regional topographic relations caused the snow to drain away in the

[1] These factors are explained in Chapter 4.

form of glaciers very much more rapidly on the western side, and to suffer great terminal losses. On the eastern side the gentler slopes induced slower glacier flow, and terminal losses were less because calving was not a large factor in the wastage at the glacier margin. In consequence the east-flowing glaciers grew and coalesced on the Bothnian lowlands into a piedmont glacier. As accumulation continued this piedmont expanded, becoming gradually broader and thicker. It enlarged until its upper surface covered the lower parts and perhaps also most of the higher parts of the mountain crest. But there appear to have been nunataks standing above the ice both on the crest and on some of the rock spurs between the seaward-flowing glaciers.

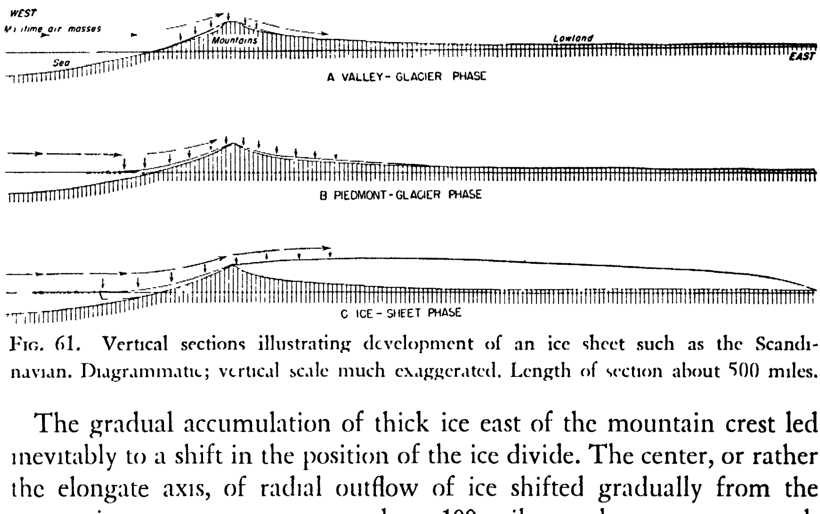


FIG. 61. Vertical sections illustrating development of an ice sheet such as the Scandinavian. Diagrammatic; vertical scale much exaggerated. Length of section about 500 miles.

The gradual accumulation of thick ice east of the mountain crest led inevitably to a shift in the position of the ice divide. The center, or rather the elongate axis, of radial outflow of ice shifted gradually from the mountain crest to a position at least 100 miles to the east, over much lower land[2] (Fig. 61). This shift has been proved by striations and other erosional features and by the distribution of erratic stones and boulders. The axis of radial outflow, as might be expected, occupied two or three clearly indicated positions east of the mountains at different times, but like the North American centers these particular positions date probably from the time of deglaciation.

Evidence from Denmark confirms the shift of the area of radial outflow. The distribution and stratigraphic relations of erratic boulders in that country show that during the Fourth Glacial age ice invaded Denmark first from the north (from southern Norway and western Sweden) and later from the northeast (from eastern Sweden and the Baltic).

In this connection it is significant that in Denmark an identical shift

[2] See A. M. Hansen 1894, p. 130; Tanner 1915, p. 698; 1937, pp. 6–7; Enquist 1916; Frödin 1925.

has been shown to have occurred in each of the two preceding glacial ages, with a striking addition. Within the later of these two drift sheets the east Swedish-Baltic element is in turn overlain by west Swedish-Norwegian drift. This arrangement means that as the Baltic ice retired from Denmark its place was occupied by ice centering in the mountains.[3] The fact that the Danish record is essentially the same in three glacial stages demonstrates that the Scandinavian Ice Sheet formed repeatedly in the same manner, although at the end of the Fourth Glacial age it may have disappeared so quickly that flow from the mountains immediately to the north did not have time to re-establish itself over Denmark.

Further confirmation of eastward shift of the ice divide is that the domed-up lake and marine strandlines[4] formed during and after the uncovering of Fennoscandia by the latest ice sheet are related to a center not in the mountains but on the west coast of the Gulf of Bothnia between 63° and 64° N latitude and between 17° and 19° E longitude (Plate 6; Fig. 79). This fact strongly suggests that the thickest ice lay far east of the mountains. Furthermore the center is elongate parallel with the mountains, suggesting that its origin is related to the presence of the mountains.

The distance of known eastward shift of the ice divide — 100 miles — added to the 75 miles or so from the mountain crest to the sea may represent the distance through which maritime air masses moving eastward were able to flow up over the steep flank of this ice sheet and precipitate snow upon it effectively.

Thickness

From the positions of erratic stones derived from the Bothnian region and transported westward across the Scandinavian Mountains, and from the amount of crustal uplift that has occurred,[5] it has been calculated that the Scandinavian Ice Sheet must have been about 10,000 feet thick in the area of radial outflow. From this maximum it tapered off strongly to the east and north and less strongly to the south. On its west flank, west of the mountain axis, it was hardly more than a series of thick coalescent outlet and piedmont glaciers.

Nourishment

Cyclonic air masses approaching from the southwest nourish the Scandinavian glaciers today and undoubtedly did so during the whole

[3] Madsen and others 1928, pp. 98, 88.
[4] See discussion in Chapter 19.
[5] See Chapter 19 and Niskanen 1939, p. 28.

of the latest glacial age. The paths followed by cyclones both now and then are shown in Plates 2 and 3. One of these paths led northeastward, skirting the Norwegian coast. The result was heavy snowfall on the entire western flank of the growing ice sheet. Another path lay along the southern margin of the ice. Because there are no high and continuous mountains in western Europe south of Norway, the air masses moving eastward along this path tended to hold the greater part of their moisture until they encountered the ice sheet. Here, along the southern margin of the ice, much snowfall occurred, aiding substantially in the extension of the glacier. This would have occurred in the spring and summer more than during the winter when frequent outbursts of cold air from the ice sheet, coupled with cold air over the ice-covered Alps, would serve as an obstacle to the moving cyclones.[6] Although such obstacles were temporary their net result would have been to give colder and drier winters to the lowlands of eastern Europe than those which now prevail.

However, as the air masses were robbed of their moisture by heavy precipitation, nourishment diminished toward the east, as is shown by the fact that the eastern part of the glacier was thin. The Timan Hills, for example, though standing only a few hundred feet above the surrounding lowlands, produced a re-entrant nearly 300 miles long in the margin of the combined Scandinavian-Siberian Ice Sheet at the time of the glacial maximum.[7]

On the north the present-day storm track that follows the Norwegian coast continues eastward along the Arctic coast of Europe and Siberia. Snowfall related to this track maintains the existing glaciers on Novaya Zemlya and perhaps also those on the islands farther north. It is probable that Atlantic air masses following this same path nourished (albeit rather scantily) the northern flank of the Scandinavian Ice Sheet.

Asymmetry

The Scandinavian Ice Sheet was strikingly asymmetric. It had a maximum southeast radius of about 800 miles,[8] reaching nearly to Moscow, whereas its west and northwest radii were never more than about 200 miles long. The explanation of this asymmetry lies in the relation of the ice sheet to the deep sea and to the movement of the moist air masses that nourished the ice. Deep water sharply limited the spreading of the

[6] Cf. Zeuner 1937, pp. 383, 387, who carried the concept to a more extreme position than the one adopted here.

[7] Similarly, during the preceding glacial age the hills between the Dnepr and Don valleys standing only 250 to 500 feet above the surrounding country, caused a re-entrant likewise nearly 300 miles long.

[8] In the Third Glacial age this radius was about 1300 miles, reaching nearly to Stalingrad.

ice sheet toward the west and north. Here the ice, though it may have formed a floating shelf of unknown width, must have broken off and floated away within a comparatively short distance beyond the present shore.

On the other hand the center of radial outflow, apparently never much more than 100 miles east of the Scandinavian Mountains and therefore near the sea, received abundant nourishment throughout its existence and thus was able to maintain a strong flow of glacier ice toward the east. The combination of this flow and the "helper process" of precipitation from maritime air masses traveling eastward along the southern border of the ice, and extending inward from the border through a considerable distance, explain the great southeastward extent of the ice sheet.

The Laurentide Ice Sheet was generated in highlands and grew westward and southward because moist air masses approaching it from the south and west nourished its windward margin. The centers of radial flow, being controlled by snowfall, necessarily migrated far to the west. On the other hand the Scandinavian Ice Sheet, also generated in highlands, grew eastward because the deep sea prevented expansion toward the west and because the Scandinavian Mountains were neither high enough nor extensive enough to create a serious rain shadow in their lee. Had they been as high and as extensive in latitude as the Rocky Mountains in North America, there would probably have been no Scandinavian Ice Sheet but instead only piedmont glaciers.

It is improbable that there were any centers of radial outflow within the eastern part of the ice sheet, because nourishment in this region was substantially less than in the Bothnian region. Indeed it is doubtful that there was any center other than the conspicuous center already described.

This simple picture of a single center revealed by simply radiating striations and eskers quite naturally influenced opinion in North America, where an analogous pattern was generally visualized for that continent. But the North American pattern was different, as has been indicated, in that the Laurentide Ice Sheet expanded *toward* the sources of moisture, whereas the Scandinavian glacier could only expand in the opposite direction.

In the North Sea region the water barrier failed to check the expansion of the ice sheet. Except for the Norwegian Depression immediately off the south coast of Norway the water is shallow, and during each of the glacial ages the North Sea floor, when not covered by glacier ice, was partly or wholly emerged. Nowhere could the water have been deep enough to float a thick ice sheet. As a result, in this sector the ice sheet spread southwest to the British Isles, a distance of 400 miles from the Scandinavian Mountains, and coalesced for a time with the much smaller British glaciers.

On its remaining side, the northeastern side, the Scandinavian Ice Sheet was coalescent at least temporarily with the smaller and thinner Siberian Ice Sheet which had spread outward from the northern Ural Mountains and Novaya Zemlya. The relations of these two ice masses are described in Chapter 17.

Extent at Maximum

The position of the margin of the Scandinavian Ice Sheet when at its maximum extent is shown in Plates 3 and 6. It can be traced in a great arc from western Denmark through northern Germany, Poland, and northern European Russia. North of the Timan Hills and in what is now the shallow southern part of the Barents Sea it coalesced with ice flowing westward from the region of Novaya Zemlya and the northern Ural Mountains. In the Barents Sea, off the coast of Norway, and west of the Shetland Islands (which it overran), it probably terminated in a more or less continuous floating shelf. On the shallow North Sea floor off northeastern England the ice sheet was confluent for a time with local ice spreading eastward from highland areas in Britain. Thus the Scandinavian Ice Sheet margin stretched across Europe on the southeast, was afloat on the northwest, and at its northeastern and southwestern ends was confluent with other large glacier masses.

The configuration of the southern margin of the ice, like that of the Laurentide Ice Sheet, was considerably influenced by the topography encountered by the glacier. This influence was especially conspicuous in Germany where the terrain reached by the ice sheet at the maxima of the Second and Third glacial ages had greater relief than that in southern Britain to the west or in western Russia to the east. From west to east across Germany the glaciated region ends at the northern flanks of highlands such as the Sauerland south of the Ruhr Valley, the Erzgebirge, the Riesengebirge, and the West Beskiden Mountains. During the Fourth Glacial age the ice sheet terminated farther north in a region of smaller relief, but, even here, according to Bülow, the limits of glaciation in places coincided with the southern ends of preglacial depressions.

The earliest Pleistocene marine deposits of eastern England (Chapter 16) include sediments derived from the southeast and believed to represent one wing of a delta of the Rhine, built at a time when part of the North Sea floor stood above the sea. Quite possibly the later blocking of the North Sea basin by the Scandinavian Ice Sheet diverted the Rhine westward and thus brought about the cutting of the valley that is now the Strait of Dover. This event would have been closely similar to the early Pleistocene integration of the Ohio River by the Laurentide Ice Sheet.

Deglaciation

Early Phases

Decline of the ice sheet, following its maximum, was marked by great and probably rapid shrinkage on its southeastern side, where it was least well nourished, and by shrinkage that was both slower and of smaller extent on its western flank, where nourishment was greatest. The deglaciation was unquestionably the result of a general rise in temperature, but it was not steady and continuous. Like the deglaciation of central North America it was marked by repeated minor re-expansions of the glacier, the outer limit of each readvance being extensively marked by end moraines. The more important of these are shown in Plate 6 and are discussed in Chapter 16.

> As the climate ameliorated the regional snowline rose and the zone of wastage widened. The output of meltwater increased; the ice thinned, and here and there [in northern Germany] was buoyed up by water ponded at its southern margin. Embayments caused by calving worked northward into the ice, facilitating the breakup of peripheral parts of the glacier into separate areas of dead ice.[9]

Thus did Grahmann describe the earlier phase of deglaciation in the southern sector of the Scandinavian Ice Sheet.

Throughout the deglaciation, but particularly during its middle and later phases, fringing water bodies, both lacustrine and marine, played an important role. In the east, lakes were ponded by the ice in the heads of valleys that normally drained northwestward into the Barents Sea, the White Sea, Lake Onega, and Lake Ladoga. In the south, in the Polish and German sectors, a whole series of small lakes fringed the ice margin. In part these were contemporaneous with the marginal "Channel River" described in Chapter 10, but some of them formed later. In time they were replaced, in the Baltic Sea basin, by a succession of lakes and arms of the sea, controlled as in North America by a combination of crustal warping and changes of outlet brought about by deglaciation.

The Baltic Water Bodies[10]

Baltic Ice Lake. After the southern margin of the shrinking ice sheet had abandoned the innermost end moraines in northern Germany and eastern Denmark, the meltwater could no longer escape directly to the North Sea region as it had been doing up to that time, for the southern

[9] Grahmann 1937*b*, p. 68. (Freely translated.)

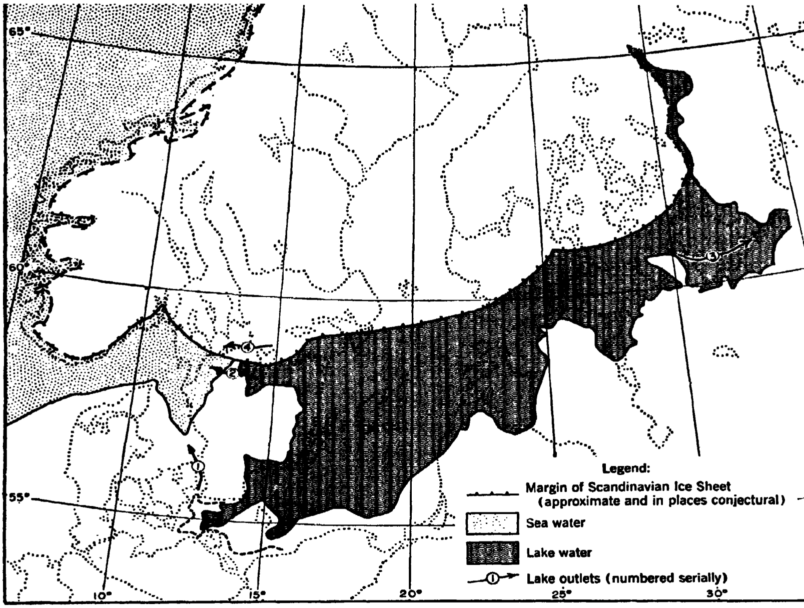
[10] Good references on various phases are Sauramo 1929; 1939; W. B. Wright 1937, p. 333.

part of the Baltic Sea region formed a shallow basin. In consequence a lake formed there, held in by rising ground on the south and east and by the margin of the glacier on the north (Fig. 62). This lake has been called the *Baltic Ice Lake* because one of its shores consisted of the ice sheet. The lake surface rose until the water was able to escape through spillway channels to the sea. According to Ramsay at least three such outlets led westward to the arm of the sea north of Denmark, and one may have led northeastward across the Lake Ladoga district to the White Sea, which then, owing to depression of the crust beneath the weight of the ice sheet, extended farther south than it does at present. Whatever the number of outlets, it is clear that they were opened successively by shrinkage of the ice sheet or closed by minor re-expansion, just as were the outlets of the earlier Great Lakes in North America. In consequence the lake level rose and fell. Thus are explained the several strandlines of the lake.

As the ice melted, the Baltic Ice Lake widened northward. When the lake attained its maximum extent the regimen of the ice sheet apparently approached equilibrium and remained in this condition for a considerable time, for a prolonged pause in the shrinkage of the glacier is marked by conspicuous end moraines lying across southern Sweden and southern Finland (Plate 6). These moraines include much deltaic material which merges northward into large eskers, thus testifying to extensive glacial drainage which discharged into the Baltic Ice Lake. Shortly after the building of these moraines was completed, deglaciation in southern Sweden exposed a new outlet so low that the lake was drained down to the level of the sea, which, because of glacial depression of the crust, covered parts of southern Norway that are now above sealevel.

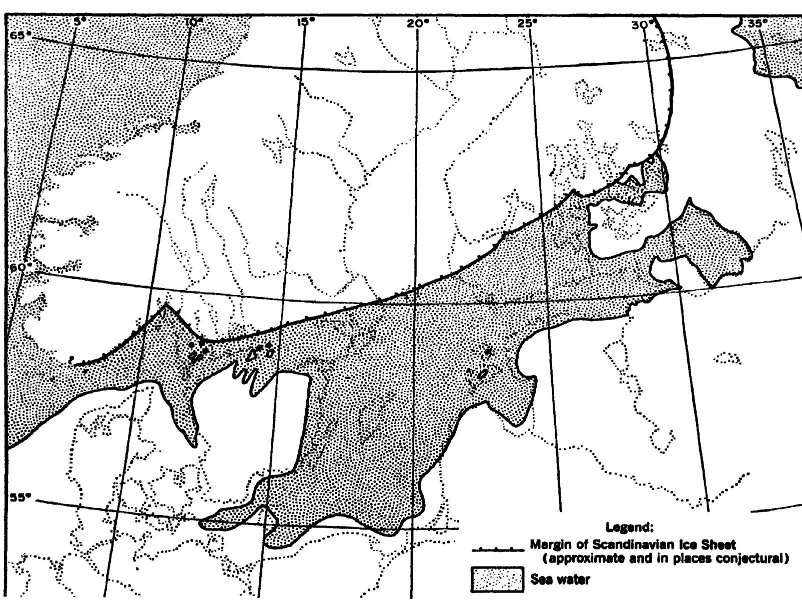
Yoldia Sea. This water body that replaced the Baltic Ice Lake is known as the Yoldia Sea (Fig. 63), named for a characteristic inhabitant of its waters, the mollusk *Yoldia arctica,* which lives today only in the colder parts of the Arctic Sea. This mollusk, and indeed the entire fauna of which it was a member, lived in the cold sea water off southern Norway while the Ice Lake occupied the Baltic basin.[11] With the draining of the lake and the entrance of the sea across southern Sweden into the Baltic basin, the *Yoldia* fauna migrated eastward. However, these mollusks flourished chiefly in the western part of the Baltic region; farther northeast the melting ice diluted the water so that its salt content was reduced to a point unfavorable for these forms of life. The Yoldia Sea is believed to have lasted in the Baltic area through several hundred years, fashioning several distinct strandlines, some of the later of which (Rhoicosphaenia

[11] A good summary of the inferred conditions is found in W. B. Wright 1937, pp 333–345.



FIGS. 62, 63. Water bodies in the Baltic Sea basin late in the Fourth Glacial age.
FIG. 62. Baltic Ice Lake near its maximum extent (*after Wilhelm Ramsay*). Dashed line shows supposed shoreline at time of outlet 1.

FIG. 63. Yoldia Sea (*after Sauramo*).



and Rhabdonema I and II strandlines, named for diatoms found typically associated with them), are strongly marked (Fig. 75).

ANCYLUS LAKE. The connection of the Baltic with the ocean, across southern Sweden, was gradually raised above sealevel as general crustal uplift[12] proceeded. This converted the Baltic water body into a lake (Fig. 76). This large body of fresh water, which has been called the Ancylus Lake after a characteristic mollusk, *Ancylus fluviatilis,* at first drained westward through a conspicuous outlet channel across Sweden. The northwestern shore of the lake consisted of the shrinking ice sheet. The lake existed for at least 1000 years, during which time the ice margin in northern Sweden shrank so far back that it ceased to be in contact with the lake, which was also receding from its western shores because of crustal uplift. Early in this time continuing uplift raised the outlet to the level of the Öresund, the channel now separating Denmark from Sweden. As the original outlet was still further raised, the discharge of the Ancylus Lake was shifted to the Öresund.

LITTORINA (TAPES) SEA. The general rise of sealevel that accompanied deglaciation gradually submerged the Öresund and converted the Ancylus Lake into an arm of the sea. There were some short-lived transitional stages, but the sea at its greatest extent is called by some the Littorina Sea and by others the Tapes Sea, for fossil mollusks found in its sediments. The marine fauna differs from the Yoldia fauna in that it records somewhat warmer water. This record is consonant with other contemporary evidence of an ameliorating climate. The extent of this sea near its maximum is shown in Fig. 77. Its history is complicated because its shorelines and floor deposits record both eustatic rise of sealevel and emergence caused by crustal uplift. Because the uplift was variable from time to time and from place to place the sea transgressed its shores in some sectors and receded from them in others. These movements have continued to the present time, and the Littorina Sea is regarded as having merged imperceptibly into the present Baltic.

Marine Transgression on the Arctic Coast

The Arctic coast of Finland and Russia is fringed, in some places widely, with fossil-bearing marine deposits that postdate the youngest till. Presumably the implied marine transgression is the counterpart of the marine water bodies that occupied the Baltic region, the deposits having been lifted, by upwarping of the crust, higher than the level reached by the worldwide rise of the sea.

[12] See Chapter 19.

Later Phases of Deglaciation

As the ice sheet decreased in thickness as well as in areal extent, topography played an increasing part in determining the direction of glacier flow. The indicator stones lodged in the till in Finland record components of flow that are significantly different from those recorded by the striations. The till was lodged on the surface at an earlier time when the ice sheet was thicker; the striations are parallel with the eskers and were made late in the process of deglaciation.[13]

Indicator stones quarried from outcrops of a conspicuous nephelite syenite on the Kola Peninsula similarly record changing directions of flow. At its maximum, the Scandinavian Ice Sheet transported this rock type southeastward but could not then carry it northeastward owing to interference by ice flowing outward from the highland of Novaya Zemlya. Later the flow from Novaya Zemlya diminished and the Scandinavian ice was able to carry the nephelite syenite northward and eastward across the Kola Peninsula to the Arctic Sea. Still later, thinning of the Scandinavian ice increased the topographic control of glacier flow to such an extent that this rock type was transported, not across the peninsula, but around it, following the White Sea basin eastward and northward.[14]

According to the reconstructions made by Sauramo from the extent of strandlines and glacial features, the Scandinavian Ice Sheet shrank inward toward the Gulf of Bothnia from the south, east, and north. Apparently also ice persisted to some extent in the higher parts of the Scandinavian Mountains. It might be supposed that the ice had disappeared from the Bothnian lowland, where it was once thickest (Fig. 61), while glaciers still occupied the Scandinavian Mountains, and indeed this seems generally to have been assumed. On the other hand the view has been put forward that in the region lying between latitudes 61° and 63° the ice remained longer east of the mountains than on the mountains themselves,[15] and it is supported by some evidence. The problem involved is similar to the relationship between the shrinking ice-sheet remnants in the Hudson Bay region and in the highlands of eastern Quebec. The true relationship is as yet simply unknown.

[13] Sauramo 1929, p. 13.
[14] Wilhelm Ramsay 1912.
[15] Mannerfelt 1945.

SEPARATE GLACIATED AREAS

British Isles[16]

Outside the limits of the Scandinavian Ice Sheet many glaciers occupied highlands in continental Europe and islands in the Atlantic Ocean and Arctic Sea, altogether covering a combined area that may have amounted to 265,000 square miles in the Fourth Glacial age and possibly more than 325,000 miles in one of the earlier glacial ages. The largest single area of coalescent glaciers outside the Scandinavian Ice Sheet was in the British Isles. Here there were no fewer than seven centers (or groups of centers) of radial outflow, each located on a highland. They are shown in Plate 6 and are listed below.

1. The great Scottish Highlands in central and northern Scotland (summits 3500 to nearly 4500 ft.). These constituted a cluster of individual centers, most of which lost their identity during a gradual merging of Scottish ice, though islands such as Skye and Mull seem to have maintained local outflow during the greater part of the time.
2. The Southern Uplands of Scotland, south of the Edinburgh-Glasgow lowland (summits exceeding 2500 ft.).
3. The Cumberland highlands ("Lake District") (summits nearly 3000 ft.).
4. The Pennine Chain (summits more than 3000 ft.).
5. The mountains of Wales, including several groups, reaching from the Bristol Channel to the Irish Sea (summits 2500 to more than 3000 ft.).
6. Mountains of Connemara and Donegal in western Ireland (summits 2000 to 2500 ft.).
7. Mountains of southern Ireland (many groups from Kerry on the west to Wicklow on the east) (summits 2000 to 2500 ft.). Although coalescent with the main British ice during one of the earlier glacial ages (the Third?), these highlands appear to have formed no fewer than ten independent, noncoalescent areas of glaciation during the Fourth Glacial age.

The altitudes of these highlands are comparatively low. On the other hand, under the prevailing maritime climate a slight reduction of temperature would result in very abundant precipitation of snow, the summer wastage of which would be greatly reduced by pronounced cloudiness. In the Scottish Highlands snow persists perennially under existing temperature conditions.

During each glacial age the British glaciers probably formed early, almost as soon as glaciers formed in the Scandinavian Mountains. The

[16] Good summaries in Kendall 1917; Slater 1929*a;* W. B. Wright 1937; all contain extensive bibliographies.

early independent glaciers enlarged and coalesced to form piedmont glaciers and small ice sheets, which themselves merged into a continuous ice cover. The coalescent ice spread westward over the Outer Hebrides and covered the continental shelf from southern Ireland nearly to the Shetlands. Probably it terminated in a floating shelf whose margin was washed by the relatively warm water of the North Atlantic Current.

In Scotland the ice grew so thick that the center of radial outflow shifted east for a time, as is attested by the transport of drift from lower country on the east, westward across highland areas. This is, in miniature, what happened in Scandinavia. But the ice thickness could not have been vastly greater than the heights of the mountains, else the centers of radial outflow would not have been as closely related to the highlands as is indicated by striations and indicator stones. Evidence of this kind shows further that the thickness of ice over the several centers and the strength of outflow from them varied from time to time, producing drift relations that are complex in the extreme. The various ice masses shouldered each other and jockeyed for position. The ice flowing westward from Scotland, England, and Wales was thicker and more powerful than the Irish ice, for it not only filled the Irish Sea basin but crowded against western Ireland, depositing drift foreign to that country.

As a combined mass the British ice came into contact with the Scandinavian Ice Sheet along the British eastern margin, from the Orkneys at the north to East Anglia at the south. Over Scotland the local ice was always thick enough to fend off the great ice sheet approaching from the northeast.[17] However, during one of the earlier glacial ages the British ice was unable to prevent the Scandinavian ice from invading eastern and central England. Even during the Third and possibly also the Fourth glacial ages, when it was everywhere less extensive and failed to reach Britain proper, the great ice sheet stood just off the British east coast, as is attested by the abrupt turning of British ice northward along the Scottish coast and southward along the English coast instead of spreading freely eastward[18] (Plate 6).

[17] This sheds some light on the probable thickness of the Scandinavian Ice Sheet. As indicated above, the Scottish ice was not thick enough to shift far away from the highlands in which it originated. Yet it was thick enough to keep the great ice sheet from Scottish shores. Hence the Scandinavian ice off Scotland can not have been greatly thicker than the Scottish ice. If we take 3500 feet as the general summit level of the Highlands and add a generous 1000 feet to the local ice, we get an altitude of 4500 feet (reckoned from present sealevel) for the Scottish ice. The Scandinavian Ice Sheet in this region could hardly have been more than a few hundred feet higher, in view of its behavior. Nevertheless we must remember that the Scottish ice doubtless robbed the Scandinavian ice of a considerable amount of moisture; hence to the north of Scotland the larger ice sheet may have thickened.

[18] The relations of the two ice masses during the Fourth Glacial age are fully discussed by Charlesworth (1931), who expressed a somewhat different opinion.

The course of deglaciation is fairly clear only for the Fourth Glacial age. The British glaciers lost contact first with the Scandinavian Ice Sheet; then with each other. They shrank, with repeated fluctuations, into the highlands in which they had originated and ultimately disappeared altogether.

The Faeroes[19]

These islands are plateaulike with a general altitude of about 1000 feet, above which mountains rise to an extreme altitude of nearly 2900 feet. Scandinavian ice did not reach the Faeroes, which were covered by a local ice sheet that undoubtedly extended outward, over the shallow sea floor around the islands, into deep water. The upper surface of the ice reached an altitude of at least 1600 feet (above present sealevel) and may have stood even higher. Striations show that the flow was radially outward. The movement should have been vigorous, for the islands are well situated to have received abundant snowfall. At some time local valley glaciers were present, instead of the ice sheet, as recorded by numerous well-developed cirques, most of them at very low altitudes.

Iceland and Jan Mayen Island[20]

Iceland, a high and mountainous island with extreme altitudes of more than 5500 feet, attained a part of its height and many of its major topographic features during the Pleistocene epoch, in consequence of faulting movements. Though much glacier ice exists today, particularly in the southern region, probably 90 per cent of the present land area was glaciated during one or more of the glacial ages (Plate 3). Striations indicate that during a late phase, at least, a composite ice sheet flowed radially out from the central part of the island, coalescing with a lesser ice sheet on the northwestern peninsula. The principal nunatak areas (marked, however, by cirque glaciers) were in the north, northwest, and northeast, that is, on the lee side of the island away from the source of snowfall. Ice thickness varied greatly, ranging from about 1200 to about 3000 feet, and probably averaging more than 2000 feet. Around much of Iceland the glacier would seem to have terminated in a floating shelf.

The evidence points to two interglacial times, but whether these had the value of stages or merely substages has not been certainly determined.

Three hundred miles northeast of Iceland lies Jan Mayen Island, a volcanic mass 30 miles long with a cone summit 7680 feet in altitude. Several glaciers descend the flanks of the cone today. The record of

[19] James Geikie 1895, p. 597; Grossman and Lomas 1895; Madsen and others 1928, p. 178.

[20] Good general references are Ahlmann and Thorarinsson 1937, pp. 168–174; Reck 1911; Thoroddsen 1905 and 1906.

former glaciation is dimmed, as in the Aleutian Islands, by extensive volcanic activity of very recent date; even a few tens of thousands of years ago the island may have had a very different form and extent, and still earlier it may not have existed at all. However, as any climatic conditions that would increase the glaciers of Greenland and Spitsbergen would also bring a glacial regime to Jan Mayen it seems virtually certain that this island, whatever its form, was glaciated during the glacial ages if it had sufficient altitude at those times.

Spitsbergen and Bear Island[21]

Glaciers cover almost 90 per cent of Spitsbergen today; the geologic evidence strongly suggests that during the glacial ages a coalescent ice mass not only covered the islands of this archipelago but flowed outward over the shallow shelf and terminated in deep water (Plate 3). Bear Island, 150 miles to the south, with a summit altitude of 1770 feet, appears to have been a local center of radial outflow of ice. This island is connected with Spitsbergen by the Spitsbergen Bank, a broad floor hardly more than 250 feet below present sealevel. During the glacial ages this bank would have formed an emergent or nearly emergent area upon which the Spitsbergen ice could have deployed, merging with the Bear Island ice.

The Alps[22]

This massive mountain arc, more than 350 miles long and with high peaks reaching 13,000 to 15,000 feet above sealevel, has received more intensive glacial study than any European area of comparable size. Many glaciers occupy its higher parts today; during the heights of the glacial ages Alpine ice covered an area that varied from 9000 to 11,000 square miles (Plate 6).

The glacial ages began with valley glaciers which enlarged until they reached thicknesses exceeding 3000 feet, overspreading the divides that separated them. Thus were formed mountain ice sheets above which only the highest peaks projected; the ice flowed down the mountain flanks, deepening the valleys, and separated into many parallel ice tongues near the mountain toe. Beyond this some of them widened to form piedmont glaciers. The southern, Mediterranean side of this great ice mass was better nourished with snow than the northern, continental side. It is probably because of this difference that during the Fourth

[21] No general reference is available

[22] The classic reference is Penck and Bruckner 1909. A better organized account is Heim 1919, pp. 197–344.

Glacial age the southern ice tongues were able to reach as far as their predecessors had in the Third Glacial age, whereas the northern tongues fell considerably short of the positions reached earlier.[23] Also attributable to this difference are the variations in the glacial flora and fauna on the two sides of the Alps. On the south the glaciers were fringed by well-watered woodland under a relatively mild climate. On the north side they discharged their meltwater into a much colder and drier country. Some of it was covered only by tundra, grazed by reindeer, woolly mammoth, and woolly rhinoceros. Analogous conditions seem to have existed in the region of the Pyrenees.

The most significant results of glacial study of the Alps have been stratigraphic, and these are summarized in Chapter 16.

Lesser Highland Areas in Continental Europe[24]

Wherever European highlands reached above the regional snowline during the glacial ages, ice formed on them. Twenty-eight glaciated areas and groups of areas are shown on the map, Plate 6. All appear to have been glaciated during the Fourth Glacial age (chiefly by valley glaciers), and in some of them direct evidence of earlier glaciations is present. In addition to the list given in Plate 6, it is possible (though not probable) that minor local glaciation may have occurred during one of the earlier glacial ages in the Erzgebirge, Thüringer Wald, and perhaps also the Harz Mountains. The evidence is so obscure that opinions in the matter differ greatly.

[23] This relationship is well shown in Leverett 1910, map plate 4.
[24] The data on these areas as a group are brought together in Antevs 1929*b*, p. 678.

Chapter 16

GLACIAL STRATIGRAPHY OF EUROPE

INTRODUCTION

The attempt to discuss the glacial stratigraphy of all Europe in a single chapter is an act of temerity, forced upon the author of this book by the necessity of providing some sort of coverage on that large and much-discussed region. Though full of interest, the task is ungrateful because it covers a field represented by a vast literature in many languages, and by complex, obscure, and conflicting stratigraphic relations in the correlation of which angels have repeatedly feared to tread. Any treatment of a vast subject by means of a summary statement is certain to imply that its interrelations are simpler than appears after detailed investigation; unfortunately this is particularly true of European glacial relations. One reason for this is that the field is complicated by the presence of a factor thus far lacking in North America except in the topmost part of the Pleistocene column: the factor of human cultural stratigraphy. This means that the same problems have been attacked by two different groups of scientists — geologists and archeologists — with two quite different backgrounds. Ideally this fact should lead to better results than as though only one group concerned itself with glacial-stratigraphic problems, and indeed in the long run it will. But not infrequently, in the past, members of one group have accepted the conclusions of members of the other group with little or no qualification and without being able to evaluate them properly. The result has sometimes been correlation involving errors that have obscured the probable truth.[1] On balance, the lack of human-cultural evidence in the greater part of the North American Pleistocene column has been an aid to the reconstruction of the stratigraphic sequence in that region, despite the greater richness of the sequence in Europe. The concerted attack by closely cooperating geologists and archeologists, now under way in Europe, should lead ultimately to general agreement upon a standard column more detailed in some respects than that which can be reconstructed for North America.

The summary that follows is based on an attempt to be objective in two distinct ways. First, as far as possible the data presented are geologic rather than archeologic. Admittedly this presentation neglects the

[1] An excellent discussion, by several authorities, of some of the difficulties involved appears in Fleure and others 1930. See also Sandford 1933.

human-cultural side and is therefore in one sense defective. But it is based on the premise that the broad outlines of North American Pleistocene stratigraphy have been developed without benefit of cultural data, yet with less complexity and disagreement than have appeared in Europe. Second, the data are presented without regard to any particular theory of the cause of the glacial-age climates. Certain theories impose rigidities on the sequence of Pleistocene climatic changes that destroy objectivity and result in attempts to fit the stratigraphic facts into a preconceived pattern rather than letting the facts determine what hypotheses are allowable.

It should now be apparent that the objectives of the present chapter are quite limited and that the discussion will be unsatisfactory to those readers who wish detailed stratigraphic information or a discussion of cultural evidence. Such readers are referred to the many excellent European references, some of which are listed among the titles at the end of this book.

The present treatment is arranged by countries. The peripheral zone of the Scandinavian Ice Sheet is considered first, then the central zone, and finally the areas of partly or wholly separate glaciation such as Britain and the Alps. This arrangement generally parallels the stratigraphic discussion of North America.

Agreement seems to be widespread (though not universal) among European geologists that there is geologic evidence of four Pleistocene glacial stages. The existence of four stages was first established in the Alps, and there has been a tendency to employ the Alpine nomenclature as a European standard. Yet to a large extent each country has its own nomenclature. The discussion that follows employs the frank expedient of referring to the four time units generally as the First, Second, Third, and Fourth glacial ages. The rigidity this terminology implies is not ideal, but it seems preferable to the frequent and elaborate circumlocutions demanded otherwise.

NORTHERN GERMANY, THE NETHERLANDS, AND DENMARK[2]

The identification and correlation of drift sheets are more difficult in Germany and Denmark than in central North America. The German drifts generally contain much sand, little clay, and little calcium carbonate (reflecting the composition of the underlying preglacial strata); hence gumbotil has not developed, and depth of leaching is less significant than in central North America. In further consequence the German tills

[2] General references are Woldstedt 1929; Wahnschaffe and Schucht 1921; Richter 1937; Madsen and others 1928. More specific review papers, each with a large bibliography, are Gams 1930; Gagel 1913; von Linstow 1913.

look less decomposed than their probable correlatives in the Great Lakes region.

In Germany the belts of outcrop are less wide and the drifts lie generally over more hilly terrain; hence they are thinner and more conspicuously eroded than the drifts of middle North America. Again, in northern Germany the streams drain north toward the glaciated region rather than south away from it; as a result there is much less outwash, by which early glaciations could be recognized, than in the Mississippi Valley region, or in the Alps or southern Russia.

Because of these factors the glacial section in the region south of the Baltic is less clear and less well understood than the corresponding section in the region south of the Great Lakes. The whole picture is complex, though undoubtedly the actual events were much more so, for as yet we see only a small piece of the record. Probably each of the earlier ages involved as complicated a sequence of expansions and shrinkages as those inferred from the Fourth Glacial stage.

In the region of northern Germany, the Netherlands, and Denmark three glacial stages have been definitely recognized, and two fossil-bearing interglacial horizons, intercalated between till sheets, are well established, being known from exposures and well records in a large number of localities. The difficulty is that the correlations of many of the interglacial horizons are uncertain.

It is generally agreed that the three recognized glacial stages are equivalent respectively to the last three of the four stages recognized in the Alps. The terminology, however, is in a somewhat confused transitional state.

Four names of glacial stages, introduced by Keilhack, have come into general use. From oldest to youngest they are Elster, Saale, Warthe, and Weichsel — names taken from German rivers in the type areas of the several drift sheets. However, opinion has turned increasingly to the view that both Warthe and Weichsel drifts are referable to a single glacial stage, the fourth or last of the Alpine sequence. Despite this general tendency, two different opinions still exist: (1) that the Warthe drift is a separate glacial stage and (2) that it is a part of the Saale Glacial stage. The position is thus very similar to that in central North America about 1930, with the Iowan drift playing the role of the Warthe.

Possible Pre-Elster Drift

That a pre-Elster Scandinavian Ice Sheet existed there can be little doubt, but like the Nebraskan drift in eastern North America it has not been recognized with certainty. Some evidence of such an early glacia-

tion is found in the Baltic coastal region, in "late Tertiary" strata that contain partly decomposed erratics of northern Scandinavian origin. The evidence is very obscure, and the interpretation can only be regarded as doubtful.[3]

Tegelen Horizon[4]

More definitely pre-Elster and almost certainly interglacial are deposits occurring at many localities along the lower Rhine and taking their name from the Dutch village of Tegelen. It is generally believed, despite contrary suggestions, that all the exposures belong to a single horizon. The material consists of sand, silt, and clay, with peat containing elm, ash, maple, hornbeam and grape, shells of the extinct fresh-water snail *Paludina diluviana* and other mollusks, and bones of the elephants *Loxodonta (Hesperoloxodon) antiquus, Mammuthus (Parelephas) trogontherii,* and *Mammuthus primigenius,* two species of rhinoceros, hippopotamus, deer, horse *(Equus stenonis),* and hyaena. This assemblage points in general to a temperate climate.

The Tegelen horizon overlies terraced fluvial deposits thought to represent the Rhine valley train of a pre-Elster glaciation and underlies a younger Rhine outwash that may be Elster. The stratigraphic position of the Tegelen interglacial, therefore, though not fixed definitely, is apparently pre-Elster.

Elster Glacial Stage

The Elster stage is the most extensive of the drift sheets in Germany, and in Poland as well, for though its area of outcrop is not great it extends beyond the younger drifts across Poland and westward across Germany at least as far west as the upper Weser River. Farther west it is uncertain whether the outermost drift is the Elster or the next younger drift sheet.

The Elster drift consists of till, comparatively thin in many places but locally reaching 100 feet, erratic stones and boulders, and lacustrine deposits. The till is deeply oxidized and leached and is so extensively eroded that in some districts it consists of isolated separate patches. Its surface expression is erosional rather than morainic; it is without undrained depressions. In some places the till has been eroded to form a residual boulder pavement, sharply overlain by Saale drift. In some districts there is no till at all; the drift is represented merely by scattered erratics of Scandinavian origin.

The lake deposits occur in entrants along the northern flanks of highland areas. Evidently they represent ice-marginal lakes formed where

[3] The case is summarized in Richter 1937, pp. 85–92; see also pp. 97, 98.
[4] The basic reference is Eugène Dubois 1904.

the ice sheet blocked drainage that normally flowed northward. Zones of intense crumpling and faulting in the lake sediments seem to indicate episodes of temporary advance of the glacier margin, which thrust forward over the lake floors. Two and possibly three conspicuous re-expansions of the ice sheet have been inferred from evidence of this kind. The Elster drift is mantled with loess.

Elster-Saale Interglacial Stage

Several fossil-bearing interglacial deposits are generally assigned to the interglacial age that followed the Elster Glacial age, though the stratigraphic position of some of them is not entirely certain. Near Lauenburg on the lower Elbe River a well-developed peat bed overlies till thought to be Elster and is overlain by outwash correlated with the Saale glaciation. The peat records a considerable length of time during which a pine-birch forest gradually gave way to an oak-ash assemblage, implying a change from cool climate to temperate climate.[5]

In places the Lauenburg peat is unconformably overlain by marine silt containing a cool-water (though not Arctic) mollusk fauna. The implied water body, an arm of the North Sea, was called by Penck the Holstein Sea. As the importance of the unconformity separating the marine deposits from the underlying peat has not been evaluated, it is possible that the two horizons are parts of different interglacial stages.

More definitely placed stratigraphically are the *Paludina* deposits in the Berlin district and the interglacial deposits of the Luneburg Heath, each of which lies between two tills that are fairly certainly Elster and Saale. The Berlin deposits consist of fine alluvium with fresh-water mollusks, distinguished by *Paludina diluviana;* also fishes, water plants, land snails, and the alder. The implied climate is similar to that of the present day.[6]

The deposits in the Luneburg Heath region, south of Hamburg, are exposed at many places. One exposure, near Ülzen,[7] is a marl intercalated between two tills. From it have been taken fossil fishes, pollen of fir, birch, and pine, and the bones of rhinoceros, bison or ox, and deer. Similar deposits exposed nearby contain oak, hazel, alder, and poplar. Near Wiechel in the same region is another fresh-water deposit underlying till correlated with the Saale stage. The deposit contains fishes and diatoms as well as fossil fir, pine, oak, beech, maple, linden, and holly. Both trees and diatoms testify to a mild climate.

[5] Schlunk 1914.
[6] Schmierer 1922.
[7] Jessen and Milthers 1928.

In Denmark deposits at a number of localities combine to record a part of the Elster-Saale Interglacial. At the base of the composite section is a marine clay (Esbjerg deposits) with an Arctic mollusk fauna characterized by *Yoldia arctica*. Above this are the Vognsbøl marine deposits recording a gradually rising water temperature. (It is supposed that the sediments of the Holstein Sea, already mentioned, are next in the regional succession.) Also assigned to the Second Interglacial are several lacustrine deposits consisting of marl and diatomaceous earth. Contained fossil plants show a transition from northern coniferous forest to temperate (oak) forest, which in turn was replaced by northern conifers. Probably this section represents the earlier part of the interglacial, the record of the later part having been lost.[8]

Saale Glacial Stage

In Germany the Saale drift is generally less extensive than the Elster, though in European Russia it is more so, and in places it has not been clearly differentiated from the Elster. It has been considerably eroded, and in some regions it consists only of patches. The effects of oxidation and leaching are conspicuous, but the actual depths of alteration are of doubtful significance, owing to the great and variable permeability of the material. Erosion is not so extensive but that the till retains some morainic topography; in Russia as well as in Germany subdued end moraines can be traced through considerable distances; it has even been thought possible to identify two or more substages within the Saale stage. In places large masses of outwash and areas of kame-like ice-contact stratified drift are present. Stratigraphic sections near Leipzig record three distinct tills, all referable to the Saale, separated by lake and stream deposits. This drift sheet is generally covered with loess which, however, in some places was deposited only after deep oxidation of the underlying drift. Drift of this age is also believed to be present in Denmark.

Saale-Warthe Interglacial Stage

To this stage probably belong the Eem deposits,[9] which antedate the Brandenburg moraine and postdate the Saale drift. Their relation to the Warthe drift is a matter of debate, partly because they are not everywhere in place, some of the best exposures being in large masses transported and contorted by a succeeding ice sheet. The Eem deposits are really a

[8] See an excellent table in Madsen and others 1928, p. 94.

[9] The basic reference is Madsen and others 1908; amplified by Nordmann 1928. See also Madsen and others 1928; Woldstedt 1929, p. 139.

zone or horizon, chiefly marine and with distinctive fossils, included within a longer sequence of inorganic lake sediments and peats, a sequence to which the names Brörup and Herning are applied in Denmark. The name is derived from the small river Eem, a tributary to the Zuider Zee in the Netherlands, and the deposits are widespread, having been found at many localities in Belgium, the Netherlands, Denmark, and northern Germany as far east as East Prussia.

The physical character of the Eem sediments records progressive submergence followed by emergence. The contained fauna is distinctive, testifying to water as warm as or warmer than that in the southern part of the North Sea today. This fauna constitutes a strong argument in favor of complete deglaciation and thereby strengthens the assignment of the Eem zone to the Saale-Warthe, an interglacial apparently having the value of a stage, rather than to the Warthe-Brandenburg interval, which has only the value of a substage. Near the top of the Eem sequence the fauna records a decrease in water temperature and by inference the approach of another ice sheet.

A boring at Skaerumhede in extreme northern Denmark has revealed a sequence of marine sediments 400 feet thick that may be slightly younger than the Eem zone.[10] Its rich content of fossil mollusks demonstrates a gradual transition from a climate not unlike the present one, to an Arctic climate, and hence by inference the oncoming of a new glacial age, apparently the Fourth. Unfortunately the only certainty about its stratigraphic position is that it is pre-Brandenburg.

Peat exposures in various parts of Denmark have been combined to give a picture of the changing plant cover and fluctuating temperature conditions during this interglacial age.[11] The climate fluctuated between temperate and subarctic.

Fourth Glacial Stage

Because of the uncertainty referred to above, as to the correlation of the Warthe drift, no general name appears to have been given to the Fourth or latest glacial stage in northern Germany and the regions adjacent to it. The name Weichsel, used by Keilhack in this sense, excluded the Warthe drift which nevertheless is now thought of as the earliest substage of the latest glacial age. Accordingly it seems best, in the present discussion, to designate the latest glacial stage as the Fourth, and to avoid as far as possible the use of the term Weichsel, which, until the nomenclature has been cleared up, would only lead to ambiguity.

[10] The basic reference is Jessen and others 1910.

[11] Madsen and others 1928, p. 106.

Warthe Glacial Substage

The position that the Warthe drift is an early substage of the Fourth Glacial stage is not unassailable, but it accords with the trend of recent opinion.[12] This drift forms a belt of outcrop extending from the North Sea coast of Germany (where it is narrow) eastward across middle Poland (where it is wide) into western European Russia, east of which it has not yet been mapped. It is characterized by constructional, morainic topography with some closed depressions (its outer limit in northwestern Germany is marked by the Fläming end moraine) but throughout much of its extent the topography has been sufficiently modified by mass-wasting so that it is subdued as compared with the succeeding Brandenburg drift. In degree of modification, in fact, the Warthe and Iowan drifts seem comparable. The fact that the Warthe drift is related to the "lower terrace" of the Rhine, whereas the Saale drift is related to the "middle terrace," indicates a long elapsed time between these two drifts. The Warthe drift sheet is covered in places with loess.

Warthe-Brandenburg Interval

The time that elapsed between the Warthe and Brandenburg glacial maxima is indicated in part by the difference in degree of erosion exhibited by these two drift sheets and in part by fossil-bearing stratified deposits. These deposits are known from many exposures and well records in northern Germany (chiefly in the vicinity of Berlin) and in western Poland. In most places they lie between two tills that are confidently referred to the Brandenburg and the Warthe substages respectively. An outstanding representative section is the Rixdorf sequence, exposed in a gravel pit at Neukölln (formerly Rixdorf), a suburb of Berlin. The name Rixdorf is applied to the whole group of deposits of this date in the Berlin region. The rich mammalian fauna collected (not all at one locality) from this horizon includes groups suggesting some what different climatic conditions. Mammoth, woolly rhinoceros, musk-ox, reindeer, and Arctic fox suggest a boreal climate; two other elephants, *Rhinoceros merckii,* lion, hyena, beaver, deer, bison, ox, moose, and horse suggest varying degrees of warmer climate. The assemblage has been interpreted as two faunas, one warm and a later colder fauna immediately preceding the approach of the Brandenburg ice. The bones of certain elephants and other animals are present but seem to have been reworked from much older deposits. In fact it seems probable that the relations are more complicated than has been realized and that further

[12] See summary in Gams 1938.

study will be necessary before a satisfactory climatic interpretation can be arrived at.

Brandenburg (German) Glacial Substage

A belt of country in northeastern Germany, including Berlin, is underlain at the surface by a till sheet ending on the south in a conspicuous end moraine, the Brandenburg moraine. This is the drift sheet that has been called the Weichsel, but because the name Weichsel has been used with different meanings it seems best for the present to identify this drift by the name of its end moraine, concerning the limits of which there is little question.[13] The drift is widely regarded as the result, not of a pause in the shrinkage of a formerly more extensive ice sheet, but of a marked re-expansion following an earlier shrinkage of unknown amount. For this reason it has been thought of as a true substage.

The Brandenburg substage consists of till and stratified drift. It is the oldest of the drifts that have fresh surface expression with conspicuous morainic topography, the details of which, including an abundance of closed depressions, are fresh and little altered by mass-wasting and other processes of erosion. It has this character in Poland and European Russia as well as in Germany, although in Russia the drift is thinly mantled with loess, whereas in Germany it is not. Apparently the difference results from the fact that the climate becomes drier toward the east. The relations are analogous to those in North America, where Wisconsin drifts that are loess covered in the Central Region have little or no loess in the East.

Scattered end moraines, not conspicuously developed, occur as a part of this drift sheet north of the Brandenburg moraine itself, indicating minor pauses in the recession of the ice-sheet margin if not actual readvances. In addition to these, a more conspicuous end moraine stretches across northern Germany and Poland, passing north of Berlin and overlapping the Brandenburg moraine both in western Poland and in northwestern Germany. It is known as the Outer Baltic (Frankfurt) moraine in Germany and the Poznań moraine in Poland. There can be little doubt that this line marks the limit of a re-expansion following a shrinkage, but the relative importance of the re-expansion is not altogether certain. It is not generally regarded as having substage value.

During this substage and the one following were fashioned most of the *Urstromtäler*, in part from pre-existing valleys and other depressions.

[13] It should be emphasized that the term *Brandenburg Glacial substage* has no official stratigraphic standing. It is used here only to clarify the relations as we understand them and pending the general adoption of an improved terminology. The correlative term *German substage*, used by Daly (1934, p. 54), has merit.

The outflow of meltwater through these channels constituted an important element in the progress of deglaciation at this time.

All of Denmark except the western part was then covered by the ice sheet. The drift border is distinct, but whether it correlates with the Brandenburg or the Frankfurt-Poznań border is apparently unsettled.

Pomeranian (Danish) Substage

The next conspicuous re-expansion of the Scandinavian glacier, following a shrinkage, is marked by a drift sheet having at its southern limit a conspicuous end moraine, the Pomeranian, Inner Baltic, or Great Baltic moraine. This great ridge of drift extends with few interruptions from northeastern Denmark around the southern Baltic region into Lithuania, beyond which it has not been traced. The substage represented has been called the Danish substage.[14] Its outer limit in Denmark is the "East Jylland Advance" of Danish geologists.

Repeated minor expansion and shrinkage occurred during the general deglaciation that began with the abandonment of the Pomeranian moraine as indicated by several end moraines.[15] One such fluctuation is recorded in East Prussia by lake deposits lying between two tills and containing a mollusk fauna. The climatic implication of the mollusks is equivocal; consequently the extent of deglaciation at this time is not known, although it was not necessarily great.[16] The chief re-expansion of the ice sheet at about this time, marked by a distinct end moraine in southeastern Denmark, is known to Danish geologists as the "Belt advance."

The contemporary deglaciation in the region of the Danish Islands was complicated by the flow of ice from two sources, the Baltic region and the mountains in southern Norway—flow that was not synchronous. In consequence the recorded ice margins are strongly lobate and the local stratigraphic relations are complex.

Scanian Glacial Substage

Repeated oscillations of the ice-sheet margin eventually led to a re-expansion that produced an important end moraine across the southern part of Scania in southern Sweden and across the extreme southeastern part of Denmark. The substage represented by this moraine and the deposits made during the subsequent shrinkage have been called the

[14] By Daly (1934, p. 54) in part following De Geer (1912, pl. 1).
[15] Shown in generalized form in Richter 1937, p. 135.
[16] Richter 1937, p. 150.

Scanian substage.[17] In Denmark the related ice-sheet expansion is known as the "Langeland advance." East of the Baltic the correlative features have not yet been differentiated.

The Baltic Ice Lake, described in Chapter 15, had come into existence. One of the temporary retreats of the glacier margin that occurred during the general Scanian deglaciation is recorded at many places in eastern Denmark and also probably in southern Scania by exposures showing two horizons containing Arctic plants, separated by a distinct horizon of silt or peat containing plants, fresh-water mollusks, and mammals of subarctic character. The warmer-climate zone is known as the Allerød phase.[18]

Bothnian Glacial Substage

The final conspicuous substage is represented by massive end moraines that cross the Oslo Fjord district in Norway (the Ra moraines), the lake region of central Sweden, and the southern part of Finland (the Salpausselkä moraines).[19] These and the deposits made during the subsequent retreat have been grouped together as the Bothnian substage.[20] The moraines do not constitute a simple line. In some sectors there are two of them, and in others there are several minor ridges. In Finland there are two (and in places three) moraines, close together, each in part deltaic, representing deposition in the Baltic Ice Lake. This water body came to an end shortly after the beginning of this sub-age and was replaced by the sea.

There is little evidence of any further pronounced re-expansion of the waning ice sheet after the abandonment of the Swedish and Finnish moraines. Accordingly the Bothnian is the last substage of the Fourth Glacial stage. A distinction between "glacial" and "postglacial" has been drawn on a frankly arbitrary basis, but as stated in Chapter 11 there are logical grounds for avoiding such usage as being artificial and more misleading than helpful.

Summary

Summarizing, we have the tentative stratigraphic column for northern Germany and the regions adjacent to it, shown in Table 10.

It is worth pointing out that the substages of the Fourth Glacial are in Europe identified mainly on the basis of conspicuous end moraines

[17] Daly 1934, p. 54.

[18] Summary in Madsen and others 1928, p. 133; see also Antevs 1928*a*, p. 158.

[19] In Finland this substage has been defined as beginning along a line somewhat north of the moraines, because the moraines in Sweden and Finland may not be strictly contemporaneous. See compilation in Antevs 1928*a*, p. 165.

[20] Daly 1934, p. 56. By De Geer (1912) the corresponding time unit was termed the "Finiglacial Subepoch."

TABLE 10. TENTATIVE STRATIGRAPHIC SECTION FOR GERMANY AND ADJACENT REGIONS
(Compare Table 28)

Northern Germany and the Baltic Region Generally		*Denmark*		*Sweden* (DeGeer 1925)
Stages	Substages	Substages	Stages	
Fourth Glacial	Bothnian Scanian Pomeranian (Danish) Brandenburg (German) *Warthe-Branden-burg interval* Warthe	Langeland Belt East Jylland (First advance) (Not clearly recognized)	Third Glacial (in Denmark)	Postglacial Finiglacial Gothiglacial Daniglacial
Saale-Warthe Interglacial			*Second Interglacial*	
Saale Glacial			Second Glacial	
Elster-Saale Interglacial			*First Interglacial*	
Elster Glacial			First Glacial	
Pre-Elster Glacial?				

and stone counts, whereas in North America they are differentiated more commonly on a basis of degree of decomposition of the drift sheets. In both regions, however, both lines of evidence have been used.

Whether any of the substages outlined above are the exact correlatives of Wisconsin substages in North America is debatable. It has been suggested that some correspondences exist,[21] and indeed they may, though they seem hardly striking enough to be compelling. When we remember that even along the southern margin of the Laurentide Ice Sheet the history of deglaciation differed in different sectors, we are hardly justified in proceeding without the utmost caution. The most

[21] Cf. Bryan and Ray 1940, p. 67.

attractive correspondence is perhaps that suggested between the Iowan drift and the Warthe drift, yet on those two drifts we have comparatively little information.

POLAND[22]

Most of the stratigraphic units described as occurring in northern Germany are present also in Poland. Stratigraphic terminology in Poland varies. Both the north German and Alpine stratigraphic names are used, and in addition there exists an entirely Polish nomenclature, which is as follows:

STAGES (*include some substages*)	APPARENT GERMAN EQUIVALENTS
Varsovien II Glacial	Brandenburg (or Frankfurt-Poznań?) Glacial substage
Masovien II Interglacial (Deposits at Zoliborz, Timoszkowice, Zdowszczyzna, etc.)	Warthe-Brandenburg interval
Varsovien I Glacial	Warthe Glacial substage
Masovien I Interglacial (Deposits at Szczercow; deposits with *Paludina* near Warsaw)	Saale-Warthe Interglacial stage
Cracovien Glacial	Saale Glacial stage
Sandomirien Interglacial	Elster-Saale Interglacial stage
Jaroslavien Glacial[23]	Elster Glacial stage

Each of the interglacials named in the Polish column is well represented by fossil-bearing deposits, as suggested in the table. The Sandomirien deposits record a nearly complete interglacial plant sequence grading from subarctic coniferous forest at their base, upward through pine-birch-hazel-oak forest, and showing a transition from cold to temperate climate. Apparently, however, the Sandomirien does not represent a pre-Elster Interglacial, as some reviewers have mistakenly supposed, but rather the Elster-Saale Interglacial.

EUROPEAN RUSSIA[24]

In European Russia three glacial stages have been recognized with confidence and a fourth (earliest) stage is believed to be represented in addition. As in Poland, there is a local nomenclature, but the Alpine terminology is also used. The available information is briefly summarized in outline form:

FIRST GLACIAL STAGE. Possibly, represented by the lowermost till in an

[22] Summary in Klimaszewski 1932; see also Szafer 1931; Woldstedt 1933.

[23] Not represented by drift, but inferred from subarctic plant remains in the base of the Sandomirien deposits.

[24] General references are Gromov 1945; Spreitzer 1941; Gerasimov and Markov 1939; Woldstedt 1933; von Bubnoff 1930.

exposure at Likhvin on the Oka River. The First Interglacial is not yet certainly represented.

Second Glacial stage. (= *Likhvin* = *Elster Stage*). Less extensive than the succeeding stage; the position of its border still debatable. Represented by till recognized in many exposures and well records. The till exhibits some decomposition. This stage is represented also by outwash remnants in the valley of the middle Volga River.

Second Interglacial stage. Represented by deposits at the Likhvin locality, with a flora including fir, pine, birch, elm, beech, hazel, and several water plants, indicating a climate at least as warm as the present climate in the same district. At Odintsovo near Moscow there are exposed between tills of the Second and Third stages silt beds containing birch, alder, mammoth, musk-ox, and horse. According to Gerasimov and Markov deciduous forests then exceeded their present limits, extending east beyond the Ural Mountains and northwestward into southern Finland.

Third Glacial stage (= *Dnepr* = *Saale Stage*). The most extensive drift sheet in European Russia, forming the great Dnepr and Don lobes (Plate 6). This is in contrast to its extent in Poland and central Germany, where it fails to reach the outer limit of the Elster glaciation. This relationship is not unlike that in central North America, where the Illinoian drift was most extensive in Ohio, Indiana, and Illinois, whereas farther west, in Missouri and Iowa, the Kansan drift was more extensive. Probably in both regions the difference is basically the result of climatic differences, in part induced by the presence of the ice sheets themselves.

Discontinuous end moraines are present in this drift in European Russia, but they are neither numerous nor conspicuous. There are also considerable bodies of outwash. At Putivl, northwest of Kharkov, in the area of the Dnepr lobe, *Mammuthus primigenius* occurs at the base of the drift.

Third Interglacial stage. The status of the Warthe drift appears to be even less certain in Russia than it is in Germany. However, recurrent exposures of a distinct peat horizon, overlying the Dnepr drift in the belt of country trending from eastern Poland through Minsk and Moscow to Gor'kiy, are probably referable to the Third Interglacial. Their stratigraphic position is fixed in part by their relation to stream terraces. The peat is characterized by aquatic plants including a typical species, *Brasenia purpurea;* hence the flora is known as the Brasenia flora. A marine transgression with boreal mollusks referred to this stage is recognized in the White Sea and on the Arctic coast near the mouth of the Pechora River.

FOURTH GLACIAL STAGE. The Warthe drift continues into Russia, but both its correlation and the position of its border are uncertain. The Brandenburg (or Frankfurt-Poznań?) substage is correlated with the Valdai substage of the Russian nomenclature. At least two substages (other than the debatable Warthe) are recognized in Russia. The evidence consists of peat beds intercalated between two tills and containing Arctic or subarctic floras.[25] The positions of the borders of the substages are very imperfectly known.

WARM-WATER MARINE INTERGLACIAL. On the northeast coast of the Kola Peninsula facing the Barents Sea, and north of latitude 68°, there occur marine deposits with a warm-water molluscan fauna. This deposit is regarded as interglacial,[26] and it implies warming of the Arctic Sea and probably complete deglaciation of Scandinavia. Its correlation is unknown, but it is most likely related to the Third Interglacial.

SCANDINAVIA AND FINLAND

In the northern region of predominantly hard rocks, close to the center of radial outflow of the Scandinavian Ice Sheet, evidence of the earlier glacial and interglacial ages could hardly be abundant. At Våge, among the highest mountains in Norway, stream deposits containing *Mammuthus primigenius* are interpreted as interglacial,[27] as is a gorge[28] near latitude 60°, similar to the interglacial gorge in New York State mentioned in Chapter 14.

On the Bothnian coast of Sweden, near latitude 65° 30′, is an organic lake sediment overlain by till and containing fossil plants and insects that point to a climate like that in the same region today[29] and therefore indicate a condition of complete deglaciation. The relations of this deposit suggest that it antedates the Fourth Glacial age. A similar deposit occurs farther south in the same region, but its stratigraphic position is more obscure.

In contrast to this very small amount of information, much more is furnished by Denmark, southern Sweden, and the Oslo district in Norway. The stratigraphic section in Denmark has already been discussed. No mention has been made, however, of the marine deposits of the Fourth Glacial age exposed around the North Sea basin in Denmark and southern Norway.[30] The retiring ice-sheet margin in the region of

[25] Gromov 1945, p. 508; Spreitzer 1941, p. 20.
[26] Gerasimov and Markov 1939, p. 448.
[27] A. M. Hansen 1894, p. 128.
[28] Antevs 1929*b*, p. 662.
[29] Antevs 1929*b*, p. 659; Gerasimov and Markov 1939, p. 448.
[30] See summaries in Madsen and others 1928, pp. 163–169; W. B. Wright 1937, pp. 333–343.

northern Denmark and the Oslo Fjord in Norway was evidently bathed in sea water, for the marine deposits there are closely related to moraines and ice-contact drift features. The marine sediments contain faunas in which the mollusks *Yoldia, Arca,* and *Saxicava* are conspicuous. Deposition began before the beginning of the Pomeranian sub-age and continued throughout the existence of the Baltic Ice Lake described in Chapter 15. When that lake was drained, the sea spread from the North Sea basin into the Baltic Sea basin and formed the Yoldia Sea there.

BRITAIN[31]

The Pleistocene stratigraphy of Britain is peculiarly complicated for two reasons. First, Britain was glaciated both by the Scandinavian Ice Sheet and by local glaciers. Second, the local glaciers flowed outward not from a single center but from several, and the volumes of ice spreading from each varied from time to time. The result is complex interbedding of drift sheets derived from different sources which, fortunately, are rendered more distinct by virtue of the highly varied bedrock geology of Britain, making possible the extensive use of indicator stones.

When we add to these factors the additional complications imposed by thin drifts, scanty interglacial deposits, and the frequent presence in fossil-bearing beds of secondary fossils derived from the reworking of older horizons, we get a truly difficult overall problem.

The average thickness of the British drifts and the proportion of exposed sections revealing two or more drift sheets are certainly less than those obtainable in northern Germany. The explanation probably is that northern Germany lies in the outer, predominantly depositional zone of the Scandinavian Ice Sheet, whereas Britain, having its own glaciers, lost more by erosion than it gained by drift deposits, a net loss rendered the more conspicuous because rise of sealevel has submerged large areas of glacial deposits off the present coasts of Britain. All in all, British glacial-stratigraphic research has encountered exceptional difficulties which, thanks to persistent and detailed study, are now being overcome.

Eastern and Central England

It is not surprising that eastern and central England have yielded the most complete stratigraphic section in all Britain, for this was the part of Britain that was overrun by the edge of the Scandinavian Ice Sheet and received deposits from that glacier. Also it was the part of glaciated Britain farthest removed from the centers of accumulation and outflow

[31] For general references see Slater 1929*a;* W. B. Wright 1937; C. E. P. Brooks 1919.

of swiftly flowing, eroding British glaciers. As a result eastern and central England are likely always to furnish the standard stratigraphic section for Britain just as, for somewhat different reasons, the Central Region is likely always to do so for North America.

Composite Standard Section

Even in this most favorable region authorities are not universally agreed on a standard sequence. Consequently any composite column is likely to be defective and is certain to be overgeneralized. But despite these flaws it still has more value for the general reader than local descriptions embracing the minute detail without which competent judgments can not be formed. The brief summary that follows is offered, therefore, with realization of its inadequacies.

BASAL MARINE SEQUENCE. The glacial sequence in eastern England begins with marine strata transitional from the Pliocene. These are the so-called crags (shell-bearing calcareous marine sands), and estuarine silts and clays. They are exposed in a belt several miles wide along the east coast of England centering in Norfolk, and the upper members have been interpreted as a part of an ancient delta of the Rhine, built before the opening of the Strait of Dover. Certainly the sediments were derived from the southeast. The classification of these beds is given in Table 11, in which the figures in parentheses indicate maximum thicknesses.

The faunas clearly record warm-temperature waters gradually becoming colder until actual Arctic mollusks appear at the top. Originally the entire sequence was believed to be Pliocene, but, although the Coralline crag is still unchallenged as a Pliocene deposit, opinion has increasingly placed at least a part of the Red crag, and the younger beds, in the Pleistocene.[32] The newer opinion appears to be firmly grounded in the concept that the Pleistocene epoch can be defined only on a basis of climate, and it seems likely to be sustained. It implies that the greater part of the marine sequence described records the oncoming of a glacial age.

CROMER SEQUENCE. The uppermost unit shown in the accompanying table consists of the Cromer beds, separated from the horizon beneath them by a disconformity of unknown significance. This unit has three members, all of them terrestrial, representing conditions near the shore of an estuary. A basal clay-and-lignite member is overlain by the much-discussed "forest bed," so called because it was once thought to include tree stumps in place. Actually it consists of lignite and inorganic sedi-

[32] Cf. Boswell 1936; Zeuner 1937; Pilgrim 1944.

TABLE 11 PLIOCENE-PLEISTOCENE MARINE SEQUENCE IN EASTERN ENGLAND

Implied Climate	*Stages*	*Substages*	*Zones*	
Arctic-temperate	Sicilian	Icenian	Cromerian (= Cromer beds = "Forest bed") (30 ft.) (*terrestrial*) (*Loxodonta* [*Hesperoloxodon*] *antiquus*)	
In part Arctic			Weybournian (13 ft.)	*Macoma baltica* (*marine*)
↑ transition			Chillesfordian (20 ft.)	*Yoldia oblongoides* (*estuarine*)
			Norwich crag (180 ft.) (*marine*)	Upper Div. (*Astarte borealis*) Lower Div. (*Mactra subtruncata*)
Predominantly northern		Butleyan	Red crag (40 ft.) (*marine*)	Butleyan crag (*Cardium groenlandicum*)
↑ transition	Astian	Newbournian		Newbournian crag (*Mactra constricta*)
Warm-temperate		Waltonian		Oakley horizon (*Mactra obtruncata*) Walton horizon (*Neptunea contraria*)
Warm-temperate	Plaisancian	Gedgravian	Coralline crag (50 ft.) (*marine*)	Zone of *Mactra triangula*

ments, inclosing transported stumps of trees and the transported teeth and bones of sixty species of mammals, birds, frogs, and snakes. The mammals include the straight-tusked elephant *(Loxodonta [Hesperoloxodon] antiquus), Hippopotamus,* and two species of rhinoceros, all of them suggesting temperate or warm-temperate climate.

Overlying the "forest bed" is a sand and clay member containing many land and fresh-water mollusks, land plants, and rodents. The implied climate is similar to that of the same region at the present time.

Between the Cromer sequence and the overlying till are a marine sand with both temperate and Arctic mollusks and, above it, fine alluvium with plants characteristic of the Arctic tundra. These two members have been thought of by some as the base of the Pleistocene. Viewed in a broad way the Cromer sequence and the beds immediately overlying it seem to indicate interglacial conditions terminating with the new approach of glaciation.

SCANDINAVIAN DRIFT. The Scandinavian or North Sea drift, so called because of its content of material transported from Norway and from the North Sea floor, is present along the east coast from Durham to Suffolk and extends inland as far as Oxfordshire[33] and Warwickshire.[34] It is chiefly till, oxidized, leached, and greatly dissected. Included in it are the Norwich brickearth, the Cromer till (which directly overlies the Cromer sequence on the Norfolk coast), and the Basement clay in Yorkshire. In Durham it is overlain by typical loess. During the glacial age corresponding to this drift the Scandinavian Ice Sheet not only reached Britain but even penetrated to the very center of England. There is no clear evidence that this event happened more than once.

GREAT EROSION INTERVAL. The oxidation, leaching, and great dissection of the Scandinavian drift testify to a long interglacial interval. The extreme dissection of the till in some districts is apparently the result of rapid erosion of underlying rocks that are peculiarly weak. Near Lowestoft in Suffolk a horizon (the "Middle" glacial sands) evidently dating from this time is regarded as interglacial.

LOWER CHALKY DRIFT. Overlying the weathered and eroded Scandinavian drift is an extensive and conspicuous till rich in chalk and derived from northward along the English coast. It is variously known as the Lower Chalky, Great Chalky, and Great Eastern drift and appears to be represented in Yorkshire by the Lower Purple till. The southward transport of this drift shows that, at the time, northern British ice was not free to flow out eastward over the North Sea floor, from which it fol-

[33] Sandford 1929*a*.

[34] Tomlinson 1935.

lows that the Scandinavian Ice Sheet was close at hand. Eastern England seems to have been more extensively glaciated at this time than at any other.

It should be pointed out here that all the glacial deposits described up to now together constitute what was formerly termed the Older drift. Those described hereafter were formerly regarded as Newer drift. This twofold classification has been superseded but is frequently encountered in the literature.

Hoxne interval. The Lower Chalky drift is notably dissected, though less extensively so than its predecessor the Scandinavian drift. This fact in itself suggests an interglacial interval. But at Hoxne, in Suffolk, lake deposits with peat resting on the eroded Lower Chalky drift have yielded a sequence of fossil plants and animals that record two temperate times separated by a time of subarctic climate. A part of an interglacial interval with fluctuating climate is implied. Somewhat similar deposits at Ipswich and at Hitchin are correlated with it.

Upper Chalky drift. A younger and less extensive glaciation of eastern England is recorded by the Upper Chalky (or Little Eastern) drift,[35] which consists largely of outwash and only secondarily of till and has a patchy distribution, occurring chiefly in valleys. This glaciation is thought to have been shorter than its predecessors because it produced far less conspicuous changes. The drift is derived from the north, and again the presence of the Scandinavian Ice Sheet to the east seems to be implied.[36]

Hunstanton and Hessle drifts. Younger tills (notably those of Hunstandard section. One of these occurs at Kirmington on the Lincolnshire coastal region. Although their exact relations to each other are matters of debate, and although in Yorkshire there are two distinct tills separated by a zone of weathering, it seems probable that all are referable to a single glacial stage.

Interglacial Deposits of Uncertain Stratigraphic Position

A number of interglacial strata have not been fitted into the composite standard section. One of these occurs at Kirmington on the Lincolnshire coast. It underlies till that is believed to correlate with the Hessle drift of Yorkshire and overlies a till of undetermined date. It consists of estuarine silt with a subarctic mollusk fauna. Whether it belongs to a part of the Hoxne succession or to a younger horizon, it does not in any event record full interglacial conditions.

[35] Including the Upper Purple till in Yorkshire.

[36] See an interesting contrary view in Charlesworth 1931.

A more illuminating exposure occurs in the cliffs at Holderness, Yorkshire. It consists of stratified sediments, largely marine and largely gravel, with mollusks (mostly like those living in British waters today), seal, and walrus. In addition there is present *Corbicula fluminalis,* a fresh-water mollusk of temperate-climate habitat. Despite its anomalous elements, this deposit is classed as interglacial. It lies between tills believed to be equivalents of the Lower and Upper Chalky drifts respectively; if this is so, it too would seem to belong to the same interglacial as the Hoxne deposits.

Other Parts of Glaciated Britain[37]

Although multiple glaciation in other parts of Britain has been established, no sequence of drifts approaching that exposed in eastern England has been found elsewhere, probably owing, as already indicated, to the greater efficiency of glacial erosion in northern and western Britain. In Lancashire, Cheshire, and Wales two drifts are recognized and are correlated respectively with the Upper Chalky and Hessle drifts of eastern England. In the Carlisle district in northwestern England there are drifts that appear to correlate with the Lower Chalky, Upper Chalky, and Hessle drifts, and in addition a younger drift referred to a glacial phase known as the Scottish Readvance, involving local glaciers of Scottish origin.

In Scotland itself this late phase of readvance is recorded, especially in the eastern region. In Ireland the Older and Newer drifts of the early classification have been differentiated, and in the end moraines of the Central Plains there has been recognized a comparatively late phase of the Newer drift that may correlate with the Scottish Readvance.[38]

Nonglaciated Southern England

Both the stream valleys and the interstream areas of southern England beyond the limits of glaciation yield evidence as to the Pleistocene stratigraphic column. The valleys, notably that of the Thames, provide a record of fluvial cutting and filling, visible both in fossil-bearing deposits and in terrace forms, that is related to glaciation (as shown by the provenance of sediments and by rare actual contacts with till) and also probably to the shifting position of sealevel. The interstream areas yield exposures of solifluction earths (locally called "head" and "coombe rock") that surely record the rigorous climates of glacial ages. Like the drifts, these features locally contain important records of human cul

[37] See Boswell 1932; Movius 1940.
[38] Movius 1940; Farrington (1945) presents a different view.

tures which, though not discussed here, are a source of data of great value.

When all these features, already the object of detailed study, have been fully worked out and elucidated, they will constitute an important contribution to the complex Pleistocene sequence in Britain.

Possible Correlation with Continental Europe

A correlation of the sequence described above, with that of continental Europe, has been attempted[39] and is shown in Table 12. From the point of view of both fossils and physical features of the drift sheets this correlation appears reasonable and likely to approach the truth.

Table 12. Attempted Correlation of the Glacial Stratigraphy of Britain with That of Continental Europe

	Britain	*Continental Europe*	
"Newer drift"	Phase of Scottish valley glaciation	Scanian	Fourth Glacial stage
	Scottish Readvance	Pomeranian	
	Hunstanton drift	Brandenburg	
	Upper Chalky drift	Warthe	
"Older drift"	*Hoxne Interglacial*	*Saale-Warthe Interglacial*	
	Lower Chalky drift	Saale	
	Great dissection	*Elster-Saale Interglacial*	
	Scandinavian drift	Elster	
	Cromer sequence	*Pre-Elster Interglacial*	
	Basal marine sequence	First Glacial stage	

THE ALPS[40]

Study of the Alps region has established a stratigraphic sequence that more than any other has been adopted as a standard for Europe. This widespread adoption is a natural result of the Alps region having been the subject of more intensive research, beginning at an earlier date, than any other European glaciated region. However, the Alps region is not ideally suited as a standard, for much of it is so deeply eroded that all trace of the earlier glaciations has been wiped out, and it is separated from the main region of European glaciation. As research in the latter region catches up with research in the Alps, a more complete Pleistocene record will very likely come from some plains region such as northern Germany.

The only Alpine district in which there has been found direct evi-

[39] Zeuner 1937, p. 153 (in part after Boswell); Movius 1940. See an earlier correlation, with good discussion, in C. E. P. Brooks 1919.

[40] Basic works are Penck and Brückner 1909 (with a wealth of detail but not well arranged for ready reference): Heim 1919, pp. 197–344 (less detail but very well organized).

dence, in the form of tills and interbedded fossil-bearing interglacial deposits, of as many as three glaciations is the Tyrol. Here there is a very satisfactory sequence of three drift sheets separated by two interglacial horizons, each including fossil plants.[41] Elsewhere the glacial sequence in the Alps has been reconstructed primarily from outwash, which in some valleys grades up-valley into till. There are recognized four main bodies of outwash, of which the two older are represented only by small remnants with little topographic expression, whereas the two younger are represented by extensive terraced valley trains. Their dates are fixed partly by the degree of alteration of the sediments and partly by morphology of the deposits. At one place or another, however, each outwash body is seen to be directly related to till, and the later drifts are represented not only by till but also by conspicuous end and lateral moraines. The youngest of the drift bodies, referred to the Fourth Glacial stage, yields evidence of several distinct substages, each recording a re-expansion of the Alpine glaciers. Some of the older outwash units, too, are multiple, recording more than a single episode of fluvial aggradation. In view of the multiple character of the Fourth Glacial stage this is to be expected, suggesting that fluctuations of the Alpine glaciers occurred during those ages as well. The last two glacial stages are represented also in massive end moraines. Those of the last stage are topographically fresh, whereas those of the preceding stage have been so greatly modified by mass-wasting that they retain little morainic topography.

In one sector of the northern Alps there are present outwash remnants believed to be related to one or more glacial stages still earlier than the four stages generally recognized.[42] Although the evidence is fragmentary, it emphasizes the likelihood, mentioned in Chapter 11, that on all continents the earliest glaciation generally recognized is not the earliest one that occurred in fact.

Each of the major Alpine drift units on the northern (though not the southern) side of the Alps has some loess on it, strengthening the concept that loess was deposited, under favorable conditions, at or near the maximum of every glacial age.

The glacial stages widely recognized are named for localities at which they are well displayed. In order of decreasing age they are the Günz, Mindel, Riss, and Würm.[43] They are far better developed on the northern flank of the mountains than on the southern, where some authorities have held that no more than three glacial stages are present.

[41] von Klebelsberg 1935.

[42] Eberl 1930.

[43] As an example of age differences among these stages, Mindel outwash is deeply oxidized, Riss outwash is oxidized and thoroughly leached through 3 to 6 feet, while Wurm outwash is little oxidized and is calcareous even at the surface.

The Würm stage is the least extensive of the four. Throughout the greater part of the drift-border zone the Riss stage is the most extensive. In the northeastern sector, however, the Mindel stage has a somewhat greater extent than the Riss.

Interglacial deposits are exposed sparingly, as elsewhere, and, as elsewhere, some of them are not certainly fixed stratigraphically, so that they must be fitted into the scheme of the drifts rather than themselves lending support to the drift correlations. The best known of the interglacial deposits is the Hötting breccia,[44] consisting of indurated colluvium exposed near Innsbrück and by most authorities (though without conclusive proof) assigned to the Mindel-Riss Interglacial. The colluvium contains the remnants of at least forty-two species of plants, some of which are distinctly southern and do not now live in the Alps. The implied former mean annual temperature was warmer than the present by at least 2° C., and the implied former regional snowline was higher than the present snowline by about 1300 feet. A lignitic deposit at Leffe on the Italian flank of the Alps, with a clearly interglacial plant content plus two elephants *(Mammuthus [Archidiskodon] meridionalis* and *Loxodonta [Hesperoloxodon] antiquus)* may be of the same date as the Hötting breccia.

More definitely placed in the stratigraphic column are a group of interglacial exposures[45] of peat and lignite referred to the last (Riss-Würm) Interglacial stage. The plant and animal remains recovered at the several localities are not identical, but the plants (especially the yew) point clearly to a nonglacial climate, and the mammals (including the straight-tusked elephant *Loxodonta [Hesperoloxodon] antiquus, Rhinoceros merckii, Bos primigenius, Cervus elaphus, Alce,* and *Ursus spelaeus*) constitute a predominantly nonglacial fauna.

Lacustrine deposits in comparable stratigraphic position and with similar climatic implications occur at Pianico and Calprino on the Italian flank of the Alps.

Like the latest glacial stage in other glaciated regions, the Würm stage includes recognizable substages. Somewhat different successions have been determined in different parts of the Alps, and, as these successions do not agree, correlations between them rest on uncertain ground. Unfortunately most of the substages (and subdivisions of an even smaller order) are inferred not from fossil evidence or from relative decomposition of the drifts, but from the groupings of end moraines in the Alpine valleys. Although the volume of an end moraine relative to that of other moraines in the same valley probably is a rough meas-

[44] A comprehensive reference is Penck 1921.

[45] At Durnten, Wetzikon, Morschwil, and Uznach.

ure of the relative time interval required for building it, the moraine itself gives little indication of the extent of the deglaciation that immediately preceded its construction. For this purpose other evidence is required, evidence that is rarely at hand. There is no doubt whatever that during the Würm age deglaciation of the Alps progressed in a fluctuating manner; it is only the relative lengths and correlation of the fluctuations that are in doubt.

In the northeastern and north-central Alps Penck and Brückner[46] identified, on the basis of end-moraine groupings, a substage of maximum extent of the Würm glaciers, followed by three "stadia of retreat" (named Bühl, Gschnitz, and Daun) each probably recording some re-expansion of the glaciers during the general process of shrinkage. The positions of the end moraines of these "stadia" implied that the regional snowline at the corresponding times stood about 3000, 2000, and 1000 feet, respectively, below the snowline of the present day.[47]

In the northern Swiss Alps three substages or phases of the Würm glaciation have been recognized on a basis of moraines. In order of decreasing age and decreasing extent they are the Killwangen, Schlieren, and Zürich units.[48] These are believed to be older than the Bühl unit of Penck and Brückner.

In the Bavarian Alps the farthest extent of the Würm glaciers is marked by moraines that lie outside of overridden earlier Würm moraines,[49] and a still younger phase is recorded by moraines lying within the two earlier groups.

In view of the differences implicit in these studies, and of differences of opinion regarding the relative importance of some of the later and younger moraine groups, we are obliged to conclude that the Würm substages in the Alps are far from being generally established.

OTHER SEPARATE AREAS

Two, three, or four glacial stages have been reported from various separate areas of glaciation other than the Alps. One of the more extensively studied areas, in which four stages are recognized, is that of the Tatra Mountains in Poland.[50]

[46] Penck and Brückner 1909.

[47] Formerly two intervals of extensive deglaciation, based on fossil evidence, were believed to have occurred in this sequence (a Laufen interval between two phases of the Würm maximum, and an Achen interval just prior to the Bühl). However, it became apparent later that the evidence belonged to pre-Würm parts of the column.

[48] Cf. Heim 1919.

[49] Eberl 1930. See also Knauer 1928.

[50] Eugenjusz Romer 1929.

ICELAND

That Iceland yields evidence of at least two glacial stages has long been known.[51] Solifluction earths and deflation armor overlain by evidence of glaciation may point to the existence of two interglacials.[52] Sediments at Viđidal overlying a glaciated surface and overlain by drift contain leaves, twigs, and pollen of plants that now live at much lower altitudes, implying a former milder climate. The vertical distribution of the plants through nearly 150 feet of sediments suggests a change from cool to warmer and back to cool conditions. As there is evidence that the sediments were already indurated before the Fourth Glacial age it is likely that they belong to one of the earlier interglacials.[53] A good bibliography on interglacial features in Iceland is given by Askelsson.[54]

GENERAL CORRELATION

If the several regional stratigraphic columns are now placed side by side, the result is as in Table 13. This table can not be regarded as showing firm correlations. It shows only what are considered by the majority of specialists as a series of probable equivalences. Substages of the Fourth Glacial stage are included only to a limited degree because, with these, equivalences are especially uncertain, particularly as between different sectors of an ice sheet that may have reached maxima at somewhat different times. Certainly it would be hazardous to attempt to correlate substages across the Atlantic between Europe and North America. It is not implied that substages can not be recognized more widely, not only in the Fourth Glacial stage but in the underlying glacial stages. So little is yet known about those that have been recognized that it seems wiser for the present to confine the table to its more generalized form.[55]

The suggested American equivalences may fairly be included because they are the result of a special study made in the field by a student long familiar with the age differences in the American drifts. One non-stratigraphic line of evidence that the disappearance of the ice sheets in both Europe and North America was synchronous is on record.[56] The rate of postglacial unwarping in both regions is of the same order of magnitude, and in both regions this rate has decreased during the last 5000 years to about half its former value.

[51] Reck 1911, p. 268.
[52] N. Nielson and Noe-Nygaard 1936.
[53] Lindal 1939.
[54] Askelsson 1938.
[55] More detailed tables, though now in part outdated, are found in H. R. Gale 1931 pp. 69, 75; Osborn and Reeds 1922, p. 414.
[56] Gutenberg 1941, p. 750.

TABLE 13 SUGGESTED CORRELATION OF PLEISTOCENE STRATIGRAPHIC COLUMNS IN VARIOUS PARTS OF EUROPE, AND THEIR AMERICAN EQUIVALENTS ACCORDING TO LEVERETT
(Interglacial units are shown in *italic* type)

Britain	Denmark	N. Germany	Poland	European Russia	Alps	Europe (James Geikie 1895)	Central North America (Equivalences as in Leverett 1910)
Scottish Readvance		Pomeranian					
Hunstanton	Third Glacial	Brandenburg	Varsovien II	Valdai	Würm	Upper Turbarian Upper Forestian Lower Turbarian Lower Forestian Mecklenburgian Neudeckian Polandian	Wisconsin
			Masovien II				
Upper Chalky		Warthe	Varsovien I				
Hoxne	*Second Interglacial*		*Masovien I*		*Riss Würm*	*Helvetian*	*Sangamon*
Lower Chalky	Second Glacial	Saale	Cracovien	Dnepr	Riss	Saxonian	Illinoian*
	First Interglacial		*Sandomirien*		*Mindel, Riss*	*Norfolkian*	*Yarmouth*
Scandinavian	First Glacial	Elster	Jaroslavien	Likhvin	Mindel	Scanian	Kansan
Cromerian					*Günz, Mindel*		*Aftonian*
Basal marine sequence		Pre-Elster?			Gunz		Nebraskan

* Leverett (1910, p 341) was uncertain as to the exact equivalence of this drift with the Riss drift in the Alps.

Chapter 17

GLACIATION OUTSIDE NORTH AMERICA AND EUROPE

SIBERIA[1]

Distribution and Probable Growth of Former Glaciers

The glaciated areas of Siberia are shown in the map, Fig. 64. For some areas this map is hardly more than an approximation, inasmuch as information, particularly on the Far East, is scanty. With a few modifications the boundaries on this map follow those shown in the Soviet Atlas;[2] a more conservative interpretation that may be more accurate in some parts of the Far East has been published.[3] On any map, however, the glaciated areas of Siberia are seen to be closely correlated with high land, being either confined to mountains and plateaus or including conspicuous highlands in addition to intervening or surrounding lowlands. As in western North America, every glaciated lowland area is continuous with an adjacent glaciated highland.

As indicated in Chapter 4, almost the only areas capable of supporting glaciers in Siberia today are Arctic-maritime. The mountains of the interior, although high, receive too much heat in summer to permit the existence of more than a few small glaciers. However, a broad regional reduction in mean annual temperature would lower the regional snowline in all the highland areas. If the reduction in temperature believed to have affected North America and Europe at the onset of each of the glacial ages be assumed to have affected Siberia as well, the glaciation recorded by the geologic evidence apparently could have resulted. In the interior the glaciers rarely expanded beyond the mountains, but in areas of more maritime climate, and in Arctic regions where low summer temperatures prevailed, piedmont glaciers and even a large (though not very thick) ice sheet took form through the expansion and coalescence of bodies of highland ice.

On the hypothesis of highland origin outlined in Chapter 12 the development of the Siberian glaciers may be summarized. At the begin-

[1] A summary of data and a list of references are given in Flint and Dorsey 1945*a*. See also Obruchev 1935–1938, vol. 3, pp. 1211–1245.

[2] Gorkin and others 1937.

[3] Gerasimov and Markov 1939.

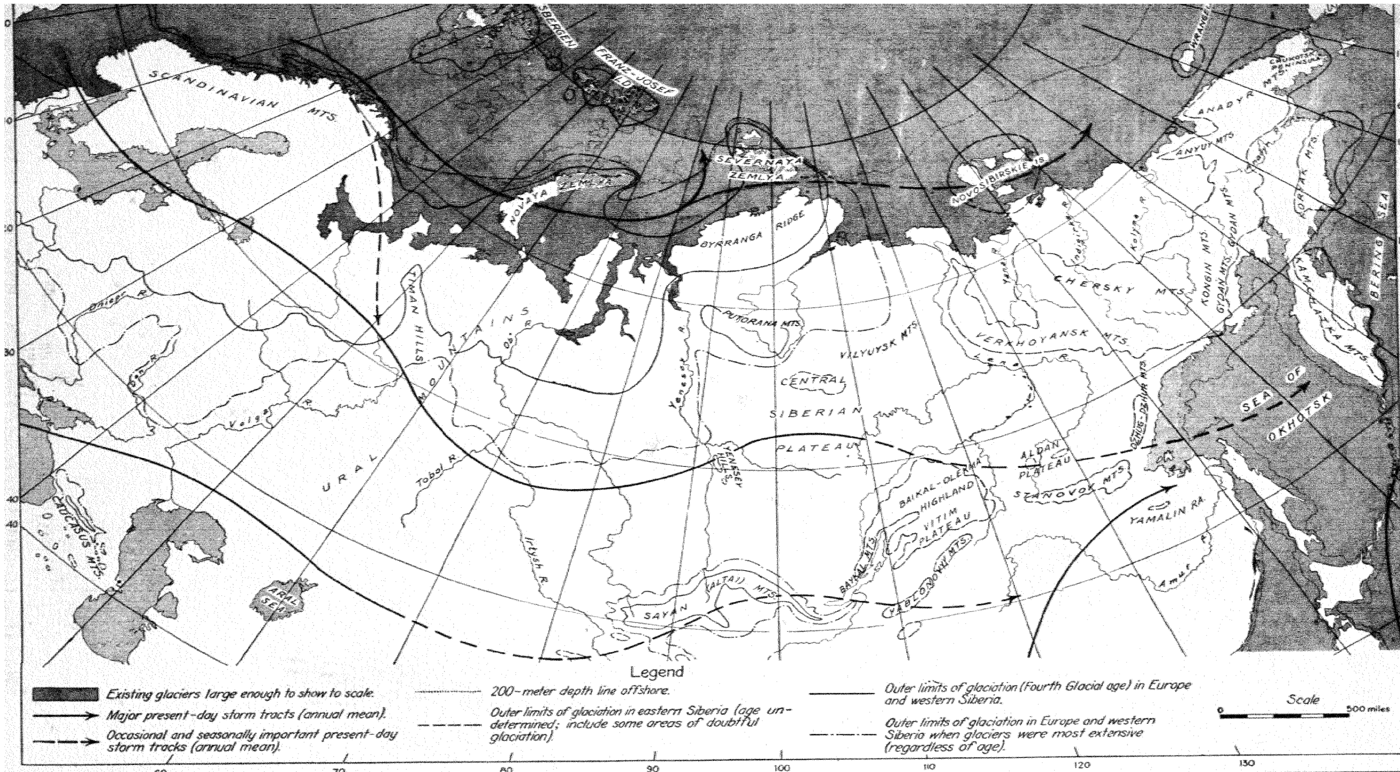


Fig. 64. Glaciated areas of Siberia. Sketch map of northern Eurasia showing existing glaciers, inferred extent of former glaciers, principal mountain units, and storm tracks.

ning of each glacial age, glaciers began to form as a result of accumulating snowfall in mountains located near the prevailing storm tracks in Europe, Siberia, and North America. Prior to the maximum of any glacial age, that is, before glaciers began to discharge extensively into the North Atlantic Ocean and Arctic Sea, the sources of moisture available for nourishing the growing glaciers of Siberia should have been much the same as the present sources (Fig. 64; Plates 2 and 3). These sources were the North Atlantic Ocean and the not-yet frozen areas of the Arctic Sea. The North Pacific Ocean supplied a more limited amount of moisture to the glaciers of extreme eastern Siberia. Despite the barrier created by the Ural Mountains, Atlantic air masses penetrate today into the interior of northern Asia (although moisture coming from the south and east is baffled out by other, higher mountain ranges); such penetration should have characterized the earlier part of any glacial age.

It seems probable that early in each glacial age, as today, cyclonic storms moving eastward from the North Atlantic region frequently became regenerated in a secondary center over the Barents Sea-Novaya Zemlya-Kara Sea area. It seems likely also that the Arctic Front in Siberia was even more active during the summer season then, because of heightened temperature contrasts, than it is today.

At first the growth of the Siberian glaciers should have been rapid under the combined influence of augmented polar-front conditions, a descending regional snowline, radiational heat losses, and migrating zones of maximum snow accumulation. As the Arctic Sea became choked with ice, Siberia was increasingly deprived of moisture from Arctic sources. With the growth of the Scandinavian Ice Sheet, Siberia must have undergone a reduction of nourishment from the much more important Atlantic sources. In consequence the Siberian Ice Sheet should have reached its maximum and should have begun slowly to shrink somewhat before the Scandinavian Ice Sheet attained its greatest extent. Still, some Atlantic air undoubtedly continued to reach the Siberian glaciers, nourishing them just enough to make the shrinkage process very slow. This condition should have continued until warming of the climate in the later part of each glacial age brought about a general decline of all the glaciers of Eurasia.

Thus it seems probable that the origin and growth of the Siberian glaciers were analogous in general principle with the development of the ice sheets in Europe and North America, but that the peculiarities of the Siberian glaciers, notably the relative thinness of at least some of them, were the result of a combination of climatic and topographic conditions peculiar to Siberia.

Northwestern Siberia: Siberian Ice Sheet

The boundary between European Russia and Siberia is the long north-south line of the Ural Mountains, nearly coinciding with the meridian of 60° E, and the northern continuation of this highland in the Arctic island of Novaya Zemlya. The Urals reach altitudes of about 5000 feet, although throughout most of their extent their crest is somewhat lower. The high northern part of Novaya Zemlya, at present covered by an ice sheet, reaches an extreme altitude of nearly 3500 feet.

East of the 90th meridian and separated from the Urals by the broad lowland of the Ob' River is a series of three more highlands: on the south the Putorana Mountains (5000 ft.); in the middle the Byrranga Ridge (3500 ft.) forming the Taymyr Peninsula; and on the north the Severnaya Zemlya archipelago (1500 ft.) in the Arctic Sea.

When glaciation was most extensive (the Third Glacial age according to recent Russian authorities) this entire group of highlands — Urals, Novaya Zemlya, Putorana, Byrranga, and Severnaya Zemlya — together with the lowlands between them, were covered by one Siberian Ice Sheet, with an area of more than 1,600,000 square miles. Measured on the flank of the Urals this ice mass was about 2300 feet thick, far thinner than the Scandinavian Ice Sheet over central Sweden. Nowhere was it quite thick enough to bury the highlands entirely.[4]

Geologic evidence indicates that glaciers formed independently in each of these highlands, spread outward, and coalesced to form the Siberian Ice Sheet. West of the Urals the ice merged with the northeastern edge of the Scandinavian Ice Sheet, which in this vicinity was thin also, because snowfall was much less here than on the Norwegian-Swedish peninsula. The Siberian Ice Sheet was fed by snowfall from moist Atlantic air masses. Initially maritime air reached the highland group by passing to the north and south of Scandinavia, bringing with it adequate moisture. However, with the growth of the Scandinavian Ice Sheet, choking of the Arctic Sea with floating ice and southward displacement of storm tracks must have reduced greatly the nourishment of the Siberian Ice Sheet. Yet some Atlantic air undoubtedly continued to reach western Siberia via the routes both north and south of Scandinavia without prior forced ascent to heights much above sea-level. The Siberian highland group now receives a greater annual snowfall than any other comparable area in the U.S.S.R.

The favorable position of the Scandinavian Ice Sheet with respect to

[4] Gerasimov and Markov (1939) believed that the glacier had no convex-up surface independent of these highlands. Whether or not this was so the genetic relation of the vast ice mass to the highlands is clear.

snowfall enabled it to expand somewhat into the area west of the Urals vacated by the Siberian Ice Sheet when, during the last glaciation, the latter began to wane.[5] This event was similar in some respects to the sequence of movements from different centers of radial outflow in the southern part of the Laurentide Ice Sheet in North America.

During the Fourth Glacial age the Siberian Ice Sheet, like its neighbor on the west, was much less extensive than during the Third. Although ice formed and flowed radially outward from each highland area, ice from the Urals and Novaya Zemlya reached southward only to latitude 64°, where its margin is marked by a conspicuous end moraine. On the east it reached the Yenesey River only near the Arctic coast, where it appears to have coalesced with the Severnaya Zemlya-Byrranga ice mass, then flowing westward. At this time the Putorana Mountains constituted a separate area of glaciation.

Central Siberian Plateau

The vast rolling region between the Yenesey River and the Lena River is the Central Siberian Plateau. Throughout most of its extent its summits reach 2000 to 2500 feet, but a broad east-west swell at the latitude of the Arctic Circle exceeds 3500 feet. This highland constituted a separate center of glaciation, which at one time expanded to some 200 miles in greatest length. Presumably this was during the Third Glacial age; whether there was ice here at all during the Fourth Glacial age is not known. Probably the ice formed a thin cap like that which covers northern Novaya Zemlya today. Cyclonic storms bringing nourishment to this ice must have come principally from the North Atlantic by the same routes as those that reached the Siberian Ice Sheet. The small quantity of ice here was undoubtedly the result of the limited amount of moisture available so far from its source.

In the southwestern part of the Central Siberian Plateau, along the eastern side of the Yenesey River valley at about latitude 60, another swell, trending north-south, is known as the Yenesey Hills. Small summit areas, 3500 feet in altitude, may have been the sites of local glaciers. If they were, probably the glaciers were contemporaneous with the greatest southeastward extent of the Siberian Ice Sheet, which at one time reached a line within 70 miles of the Yenesey Hills and induced polar-front climatic conditions in the surrounding country. It seems unlikely that this area supported glaciers during the Fourth Glacial age when the nearest margin of the Siberian Ice Sheet stood 450 miles to the northwest.

[5] Wilhelm Ramsay 1912.

Mountains of Northeastern Siberia

General Glacial Relations

Northeastern Siberia east of the 125th meridian is marked by a vast system of arcuate mountains trending in various directions. Some of the ranges are high, with Alpine summits. From west to east the principal units and their approximate extreme altitudes are (Fig. 64):

Verkhoyansk Mountains	(7000 ft.)
Chersky Mountains (several ranges)	(10,200 ft.)
Kongin Mountains	(7000 ft.)
Gydan Mountains	(7000 ft.)
Anyuy Mountains	(6000 ft.)
Anadyr Mountains	(7000 ft.)
Koryak Mountains	(6000 ft.)
Kamchatka Mountains	(8000 ft., with single peaks up to more than 15,000 ft.)

These mountains supported a complex system of valley, piedmont, and ice-cap glaciers which as a whole was much like the former Cordilleran Glacier Complex of western North America and was similar (on a much larger scale) to the system of glaciers that exists today between Mt. St. Elias and Mt. Logan in coastal Alaska.

From the mountains named and from subsidiary ranges, glaciers flowed radially outward and at their maximum extent coalesced into an almost continuous system that reached from the Lena River on the west to Bering Strait on the east, and from the Arctic Sea southward to the Sea of Okhotsk. This nearly coalescent system was 1800 miles long, almost exactly the length of the continuous part of the North American Cordilleran Glacier Complex measured from the State of Washington to western Alaska. The area involved was more than 400,000 square miles.

In addition to the main coalescent mass there were two independent areas of glaciation, the Koryak Mountains on the Bering Sea coast and the high volcanic chain on the Kamchatka Peninsula.

Throughout this entire complex little evidence of the dates of glaciation has been gathered. The glaciers had their greatest extent during a glacial age (presumably the Third) that preceded the last one, but successive glaciations have been differentiated in only a few localities. Three glacial stages are recognized in the Verkhoyansk Mountains. In general the latest glaciation, like that in northwestern Siberia, was less extensive than its predecessor, having been confined to the higher parts of the higher mountain ranges. Except in the Koryak and Kamchatka areas, the moisture available for nourishment was at all times less than that supplied to the Siberian Ice Sheet.

Maritime Region

At the head of the Sea of Okhotsk, along the Pacific coast of Kamchatka Peninsula, along the coast of Bering Sea and Bering Strait, and along the Arctic coast north of the Anadyr Mountains, the former glaciers are believed to have reached what is now sealevel, extending out over what is now the continental shelf. Both the Kamchatka and Koryak glaciated areas extend to low altitudes. Evidently these two maritime highlands received relatively heavy snowfall.

In southern Kamchatka the glaciated area reaches the sea on the east side of the peninsula, but on the west side it fails by almost 1000 feet to reach sealevel. This discrepancy suggests that during the glaciation the same difference existed that exists today between the moist eastern slopes and the less moist western slopes of the Kamchatka Peninsula.

The Arctic coast is a cold-maritime region. The glaciation of this region, centering in the Anadyr and Anyuy mountains, apparently reached the large extent claimed for it as a result less of heavy snowfall than of low summer temperatures, which reduced ablation. The present-day precipitation of this region is only one-third to one-half that on Kamchatka. The former glacier cover was probably thin as compared with the glacier ice farther south and west.

Verkhoyansk and Chersky Mountains

The massive Verkhoyansk and Chersky mountain arcs are the highest and most continuous mountain chains of northeastern Siberia. The two masses are separated by a great basin 200 miles wide, drained by the Yana River. On the outer slopes of the inclosing mountains, that is, the west slope of the Verkhoyansk Mountains and the northeast slope of the Chersky Mountains, the former glaciers were valley glaciers that hardly reached outward beyond the bases of the mountains. On the inner slopes the glaciers flowed into the Yana River basin, coalesced, and eventually filled the southern part of the basin nearly to the brim. The great accumulation of ice, 2300 feet thick, is attributable chiefly to the fact that the basin was frigid, being shielded on the west, south, and east by mountains at least 7000 feet high, so that ablation was held to a minimum. The geologic evidence indicates that at the glacial maximum virtual ice-sheet conditions were established over the entire basin, while valley glaciers continued to drain the outer slopes of the inclosing ranges. The relations thus pictured are similiar to those that seem to have characterized southern British Columbia.

Kongin and Gydan Mountains

The intermont ice-cap condition did not exist to the same extent elsewhere in northeastern Siberia because the Chersky and Verkhoyansk mountains lie farther apart than any of the other ranges. During the glacial maximum the intermont lowlands between the various ranges of the Chersky Mountains and between the Chersky, Kongin, and Gydan mountains were filled with ice which was continuous across at least the lower parts of the intervening ranges.

On their southern slopes, however, all these ranges were drained by valley glaciers. Seaward, these reached down to progressively lower altitudes in response to increased snowfall and decreased summer temperatures in the vicinity of the coast. At the south end of the Gydan Mountains the glaciers seem to have reached what is now sealevel. Likewise the eastern flank of this same mountain range was drained by valley glaciers which under the prevailing maritime climate descended to an altitude of less than 700 feet.

Chukotsk Peninsula and Arctic Coast

The glaciation of the Chukotsk Peninsula, occupied by the Anadyr Mountains, seems to have radiated outward not only from centers in those mountains but also in part from centers in other, lower mountain units, the ice from the several sources coalescing to form a covering over a large part of the peninsula.

The wide area along the Arctic coast in the region of the lower Indigirka and Kolyma rivers failed to receive a glacier cover. This was because precipitation is small and the region is so low that it does not retain winter snowfall throughout the summer. Glacier ice flowing from the high Chersky Mountains on the south did not reach this lowland because the bulk of the snowfall from moist air masses moving from the west and south was precipitated mainly on the western and southern sides of these mountains. The starved condition of this part of the Siberian Arctic coast is roughly analogous to that of the Arctic coast of Alaska, which was little glaciated because the high Brooks Range hedged off Pacific air masses, its best potential source of moisture.

It is important to repeat that the conditions described for northwestern Siberia are conditions that existed during the maximum glaciation, probably at the time of the Third Glacial age. During the latest glacial age the glaciers were much less extensive, though their limits have not yet been mapped.

Islands in the Arctic Sea

All the major island groups in the Arctic Sea — not only Spitsbergen and Franz Josef Land (discussed in Chapter 15) but also Novaya Zemlya, Severnaya Zemlya, the Novosibirskie Islands, and Wrangell Island — were completely glaciated at one or more times.

During the maximum glaciation, glaciers that radiated from Novaya Zemlya and Severnaya Zemlya coalesced with glaciers originating in centers on the mainland to form the Siberian Ice Sheet. Coalescence was made possible by the fact that the intervening water areas are shallow. The sea floor, whatever its degree of emergence may have been at that time, differed little in general from the low plains along the Arctic coast today.

Although the Novosibirskie Islands have a maximum altitude of barely 700 feet, they were the center of a small and thin ice sheet that attained a radius of more than 150 miles, as evidenced by its having expanded across the shallow sea floor and having reached the low mainland coast between the mouths of the Yana and Indigirka rivers. Apparently in this region of comparatively little snowfall nourishment was inadequate, even at the glacial maximum, to permit coalescence of this ice sheet with the glaciers that flowed northward out of the Chersky Mountains. Elsewhere than in this mainland sector the outer limits of the Novosibirskie Islands ice sheet are conjectural. Probably the DeLong Islands to the northeast were glaciated, but there is little information on them.

Wrangell Island, with a maximum altitude of 2000 feet, was the center of an ice cap similar to the Novosibirskie Islands ice cap except that it was smaller. This is known because the Wrangell Island ice failed to reach the mainland coast, though the intervening distance, across shallow water, is little more than 100 miles. Except that when at its maximum this ice sheet completely blanketed Wrangell Island, nothing is known regarding its limits. Because the position of the island is unfavorable for the receipt of snowfall, it is not likely that the ice sheet was extensive.

Highlands of the Amur-Lena Region

Dzhugdzhur Mountains

The country between the Amur and Lena rivers, east of the Lake Baykal region, includes a number of mountain masses several of the highest of which formerly generated glaciers. One of these was the Dzhugdzhur Mountains, which form the northwest coast of the Sea of Okhotsk. Although this highland reaches little more than 4000 feet

altitude, it is believed to have supported glaciers throughout a distance of more than 200 miles measured along the crest of the range. Here, evidently much orographic snow was precipitated out of moist easterly winds blowing from the Sea of Okhotsk.

Stanovoy Mountains and Adjacent Highlands

The long Stanovoy Range extends westward from the Sea of Okhotsk along the 56th parallel. Well above 4000 feet throughout most of its length, the crest in two places reaches altitudes of 7500 feet. The range is glaciated throughout a linear distance exceeding 400 miles, the glaciation reaching down to lower altitudes with increasing proximity to the sea. Although moisture from the Sea of Okhotsk supplied the principal nourishment, some cyclonic precipitation probably reached the Stanovoy Range from the southwest.

The northern base of the Stanovoy Mountains merges into the Aldan Plateau, which, near latitude 58°, rises to form two high areas respectively about 6500 and 8500 feet above sealevel. Both areas were the sites of local glaciers, fed principally by high-altitude snowfall from storms moving from the west, during the warmer months, along an intensified polar front that does not exist here today.

South of the eastern end of the Stanovoy Mountains a similar highland (the Yamalin Range), reaching an extreme altitude of 7300 feet, is believed to have carried small local glaciers. These would have received nourishment from disturbances moving eastward along the polar front, in addition to moisture from cyclonic northeasterlies blowing from the Sea of Okhotsk.

Highlands of the Baykal Region

Lake Baykal lies in the midst of a complex of high mountains and plateaus between the headwaters of the Amur and Lena river systems. The principal elements in this complex are the Baykal Mountains bordering the lake on the west, the Baykal-Olekma highland (comprising several individual ranges) northeast of the lake, and the Vitim Plateau east of the lake. All these highlands are rugged, with barren and in places serrate crests. Throughout much of their extent they stand well above 5000 feet and in places reach extreme altitudes of 9200 feet. They are the sites of a former complex of glaciers which when at their maximum extent are believed to have been coalescent. The topography of the highlands and the general character of the glaciation are like those of the Yellowstone-Teton-Wind River highland region mentioned in Chapter 12.

South of the Vitim Plateau the Yablonovyy Mountains, with a maximum altitude of 5500 feet, trend northeast. They are reported to have carried glaciers throughout a distance of nearly 500 miles. A minor highland lying south of the western end of the Yablonovyy Mountains is believed also to have been glaciated.

Probably all these highland glaciers were fed by snowfall from storms moving eastward along a well-defined polar front that existed during the summer season at the times of maximum glaciation.

Altai Highlands

The great Altai mountain system in the border region of Russia and Mongolia includes a number of ranges, of which the Sayan Range, whose rugged crest stands in most places above 7500 feet and at one point reaches nearly 11,500 feet, lies within Siberia. The Sayan was the site of a glacier complex 200 miles or more in length and in some places more than 60 miles wide, with glaciers reaching down to altitudes of less than 3000 feet. This complex resembled the former glacier systems of the Sierra Nevada and Cascade highlands in western United States. It seems probable that the Sayan glaciers derived their nourishment from cyclonic activity along a pronounced polar front that resulted from southward shift of the present Siberian Arctic Front that develops today, during the summer season, some hundreds of miles to the north.

Supposed Relation of Frozen Ground to Former Glaciation

It has been held that areas of perennially frozen ground in Siberia are complementary to the areas that were formerly glaciated. This ancient view originated in the nineteenth century but is still quoted.[6] Apparently the argument has never been stated fully and clearly, but it seems to run thus:

1. There is much perennially frozen ground in Siberia.
2. Frozen into it there have been found carcasses of the woolly mammoth *(Mammuthus primigenius)* and the woolly rhinoceros *(Coelodonta)*, which are believed to have become extinct either in late-preglacial time or at any rate no later than the Fourth Glacial age.
3. Therefore the freezing is believed to date from the Fourth Glacial age or earlier.
4. The ground could not have become frozen if it had been covered by a thick blanket of glacier ice. Hence the glaciation of Siberia must have been confined to areas not now frozen.

[6] Cf. Gerasimov and Markov 1939, p. 446. See general discussions in Nikiforoff 1928; Cressey 1939.

It is quite true that freezing of the ground has resulted from the abstraction of heat from the ground during the long and extremely cold Siberian winters. Probably it is true also that if it were covered by an ice sheet the ground would not freeze deeply, if at all. However, though the hypothesis just outlined may have sufficed when knowledge of the extent of both glaciation and frozen ground was very scanty, it does not meet the facts known today.

The areal limits of frozen ground and of former glaciation are now fairly well known. They are not complementary. In western Siberia the glaciated area extends far south of the southern limit of frozen ground, whereas southeastern Siberia, much of which was never glaciated, is widely underlain by frozen ground.

Furthermore, since no evidence has been adduced to show that mammoths and rhinoceroses did not persist into post-ice-sheet time (as mammoths clearly did in North America), the fact that the remains of these animals are found in frozen ground proves nothing regarding the date of freezing of the ground.

The growing opinion among Russian geologists is that the ground now frozen has reached that condition since the Fourth Glacial age, in direct response to the existing climate. Whether this opinion is confirmed in future, the hypothesis that, from the areal distribution of frozen ground, much can be inferred as to the extent of former glaciation in Siberia is untenable.

Multiple Glaciation[7]

Evidence of multiple glaciation in Siberia is clear and has been reported by many different geologists in localities from the Ural Mountains to the Bering Sea. In the region of the southern margin of the former Siberian Ice Sheet two glacial stages (correlated with the Third and Fourth respectively) are reported to be separated by sediments with interglacial flora and fauna. In Severnaya Zemlya two stages are recorded. In the Novosibirskie Islands there are said to be three stages separated by marine deposits. In the Verkhoyansk-Chersky mountains region there are reported at least three stages, and in the Chukotsk Peninsula two stages, of which the later is identified with the Wisconsin stage in Alaska. Three stages are recorded in the Gydan Mountains, two stages in the highlands around Lake Baykal, and at least two stages in the Sayan Mountains. In the mountain regions of eastern Siberia the evidence cited in support of multiple glaciation consists of end moraines

[7] References are cited in Flint and Dorsey 1945*a*.

and terraces rather than of fossil-bearing interglacial deposits between till sheets; its reliability must be judged accordingly.

Although correlations are still tentative, the multiple glaciation of Siberia is firmly established. Probably the several glacial ages were synchronous, though the probability is not universally admitted. It has even been held that the glaciers of dry continental Siberia could not have existed at the same time as the Scandinavian Ice Sheet in maritime Europe because the development of an ice sheet in Europe would shut Siberia off from what little precipitation it now receives.[8]

However, this view does not appear to be warranted. Early in any glacial age the preglacial storm tracks would not have been displaced southward to a significant extent, and in consequence abundant nourishment should have been available for western Siberian glaciers. Furthermore the secondary center of cyclonic activity located in the Barents Sea-Novaya Zemlya-Kara Sea area should then have been intensified, at least temporarily, by increased frontal contrasts, especially in the warmer months. Finally, although probably the Siberian anticyclone would have inhibited winter precipitation to a greater extent than at present, it should have given way to continental heating in summer, with accompanying strong cyclonic developments along the Siberian Arctic Front. Taken together these conditions should have produced a sharp increase in the total precipitation over Siberia, thus causing a relatively rapid growth of glaciers in the highland areas at the same time as in the mountains of Scandinavia.

Subsequently, as glaciers coalesced and the Siberian and Scandinavian ice sheets took form, anticyclonic circulations were locally induced over them. There is little doubt that then (but not until then) Siberia was increasingly shut off from Atlantic moisture. This deduction is consistent with the relative thinness and restricted extent of the Siberian Ice Sheet as compared with the larger and thicker Scandinavian Ice Sheet.

Thus, the belief that the Scandinavian Ice Sheet and the Siberian glaciers were not generally synchronous does not seem well founded. Furthermore it is very improbable that any climatic change causing renewed glaciation of Siberia would not at the same time reconstitute the Scandinavian Ice Sheet. Conversely, when (as at present) the temperature is too high to permit the maintenance of an ice sheet in Scandinavia, there is no possible chance of large glaciers forming in western Siberia with its high summer temperatures.

[8] Gerasimov and Markov 1939, pp. 447, 449.

Extensive Glacial Lakes and Changes of Level

Matters such as changes of level and glacial drainage in Siberia are not yet sufficiently well known to justify extensive discussion. In general, however, emerged marine deposits identified with both glacial and interglacial ages exist along both Arctic and Pacific coasts, in places extending more than 200 miles inland.[9] Undoubtedly the emergence of these deposits is the combined result of crustal warping and fluctuation of sea-level. Further, the region fringing the Siberian Ice Sheet on the south, between the Urals and the Yenesey, is said to have been (during the Third Glacial age) the site of an extraglacial lake or lakes. The impounded water, held by a broad glacier dam in the drainage basins of the Ob' and the Irtysh, seems to have been controlled by a spillway at the head of the Tobol River (lat. 65°E; long. 52°N), that discharged into a route now dry leading to the Aral "Sea" east of the Caspian.[10] Probably the development of this lake or lakes paralleled the sequences of lakes formed in Fennoscandia and in central North America during shrinkage of the ice sheets in those regions.

WESTERN ASIA

Snowfall that nourished the former glaciers of Turkey, Iran, and the Caucasus was brought chiefly by westerly winds from the Atlantic, the Mediterranean, and the Black Sea. In the Caucasus Mountains[11] the main chain (Great Kavkaz) was at one time glaciated almost continuously through a distance of nearly 400 miles. The chain is high, with many parts of the crest exceeding 14,000 feet, culminating in Mt. Elburz (18,480 ft.). In the western part of the chain the regional snowline, now at about 9000 feet, descended to 4600 feet during the maximum glaciation.[12] Eastward the snowline rose, owing to increasing aridity.

In the Little Kavkaz south of the main chain, five separate areas of glaciation have been recognized. These are on the higher parts of this chain, dominated by Mt. Alagez (13,438 ft.) and Mt. Kiambil (12,250 ft.).

In the Caucasus region two glacial stages are established, an extensive glaciation correlated with the Third Glacial stage in Europe (the Dnepr stage in Russia) and a much less extensive one correlated with the Fourth Glacial stage.

Very little is known of glaciation in Iran.[13] Kūh-I-Savalān (14,009

[9] See especially Obruchev 1935–1938, p. 1211.
[10] For a contrary opinion see Obruchev 1930, p. 275.
[11] Chief data compiled in Gorkin and others 1937, pl. 100.
[12] Gerasimov and Markov 1939, p. 449.
[13] Some recent information is available in Bobek 1937.

ft.) a volcanic cone in latitude 38°16′ N, longitude 47°50′ E, carried glaciers. The higher mountain groups[14] of the great Elburz Chain between Teheran and the Caspian Sea carried valley glaciers, some of them very long and reaching down nearly to 6000 feet.

The Zagros Mountains (dominated by Zardeh Kuh, 14,143 ft.) west of Isfahan were conspicuously glaciated. Farther northwest Takht-I-Shāh (lat. 33°20′ N, long. 49°22′ E) is so high (14,200 ft.) that it almost certainly was glaciated. Equally high mountains west, south, and east of Isfahan may also have carried glaciers, even though the regional snowline rises toward the south and east.

In Turkey some highland areas are known to have been glaciated, but information on them is very scanty.[15] These highlands appear to record former glaciation (or former greater glaciation, for some of them have glaciers at present):

Mountains of the Kurdish-Iranian border region west of Lake Urmia (11,930 ft.) and near-by Armenian Taurus (14,200 ft.).

Ararat (16,946 ft.) (39°42′ N, 44°18′ E).

Smalı Anadolu Mountains, forming the southeast coast of the Black Sea, and reaching 13,045 feet in the peak called Parhal Dağ.

Mountains in the Lake Van district in Armenia. The highest known point is Suphan Dağ (14,547 ft.), but other points exceed 13,000 feet.

Mountains between the Kara and Murat rivers, north of Harput. Several peaks exceed 11,000 feet.

Ulu Dağ (Mt. Olympus) (8179 ft.) (40°05′ N, 29°10′ E — south of Istanbul).

Erciyas Dağı (12,848 ft.) (38°33′ N, 35°28′ E — south of Kayseri).

Hasan Dağ (more than 9000 ft.) (38°10′ N, 34°12′ E — southeast of Lake Tuz).

The Toros Dağlari (Cilician Taurus Mountains) in this vicinity have summits exceeding 10,000 feet and are therefore likely to have been glaciated.

In Syria Mt. Hermon (9250 ft.) and Mt. Lebanon (10,125 ft.) carried glaciers.[16]

CENTRAL AND SOUTHEASTERN ASIA[17]

In broad view the high mountain chains of central Asia form an irregular trident facing toward the east. The long teeth of the trident are the Himalayas on the south, the K'un Lun Shan in the center, and the T'ien Shan on the north. Between the first and second is the high plateau of Tibet; between the second and third is the great Takla Makan desert basin. The three mountain chains come together in a complex knot of

[14] Notable among these are Demāvend (18,860 ft.), Takht-i-Suleimān (15,916 ft.), and Točal (11,910 ft.).

[15] See references assembled in Antevs 1929*b;* also Pfannenstiel 1940.

[16] Blanckenhorn 1914, p. 39.

[17] The literature on the glaciation of Central Asia has been well summarized in Obruchev 1930, with an extensive bibliography. See also a less specific discussion in Penck 1931.

ranges in the region in which Afghanistan, northwestern India, Sinkiang, and southern Asiatic Russia come together. The short handle of the trident extends westward into Afghanistan as the Hindu Kush and related mountains.

Northeast of the T'ien Shan is another great chain, the Altai, reaching with subsidiaries such as the Khangai and Kentei mountains into Mongolia and onward almost to Lake Baykal. East of the K'un Lun Mountains, in Sinkiang, Kansu, and eastern Tibet, are many detached high ranges separated by desert basins and plateaus. East of the Himalayas are the high ranges of southeastern Tibet, northern Burma, Sikang, and Yunnan.

All these mountain chains are made up of numerous individual ranges, some of which are well known but many of which have hardly even been explored. In consequence the map (Plate 3) showing the probable areas of glaciation is in some regions hardly more than conjectural. It is likely that the future will see greater changes in the glacial map of eastern Central Asia than in that of any other part of the world.

The ranges of the trident are high. The Himalayas, culminating in Mt. Everest (29,002 ft.) are the highest, but the T'ien Shan, K'un Lun Shan, and some of the mountains in eastern Tibet, Sikang, and Yunnan have summits that exceed 20,000 feet. Many of the higher mountains carry glaciers today. Throughout the entire region most of the higher mountains formerly supported numerous valley glaciers, many of which coalesced over the divides, though in few districts, apparently, did the coalescent masses reach the condition of mountain ice sheets.

Throughout this region the present regional snowline is very high, largely because the prevailing continental climates provide only moderate precipitation and promote ablation during the summer season. The snowline has the form of an arch or dome, highest in the driest and most continental central region, and descending to lower altitudes both on the comparatively moist Himalayas and on the comparatively cold northern ranges of the Altai. The snowline of the glacial maximum was likewise arched, rising from less than 14,000 feet in Kashmir to more than 19,000 feet in parts of Tibet, and then descending northward to 16,000 feet in the east Pamirs. Much farther north, in the Siberian Altai, it descended to less than 7000 feet.

In the Pamirs, where the three mountain chains come together, glaciation was intense.[18] Great valley glaciers up to 150 miles long were formed, and some of them coalesced to form huge piedmont glaciers at the mountain bases. The glaciation in some districts may have

[18] Nalivkin and others 1932. See also Popov 1932.

approached the mountain-ice-sheet condition. In the Pamirs three glacial stages are said to be recorded by moraine systems, and by one investigator four have been identified.

In the Himalayas glacial features record four glacial and three interglacial episodes. The glacial episodes are inferred mainly from till sheets and moraines; the interglacial from beds of alluvium and lake sediments; no fossil-bearing intertill layers seem to have been found. The last two glaciations covered barely half the territory over which the first two nearly equal glaciations spread. The first two have been correlated tentatively with the Mindel and Riss, respectively, of the Alps sequence, and the last two with the Würm.[19] If these correlations are accepted, then three glacial stages can be said to be evidenced by glacial features in the Himalayas.

On the other hand study of the alluvium, stream terraces, bedrock erosion of valleys, vertebrate fossils, and human cultures in Kashmir and the Punjab have led to the recognition of a threefold Pleistocene sequence, based both on unconformities produced by repeated mountain uplifts and on vertebrate faunas; this sequence has been correlated with the Himalayan sequence as shown in Table 14.[20] The glacials are

TABLE 14. PLEISTOCENE SEQUENCE IN THE HIMALAYAN REGION ACCORDING TO DE TERRA AND PATERSON

Stratigraphic Units		Deposits in Northern Punjab	Glacial Sequence in Kashmir
Pleistocene	Upper	Sediments Main terrace Potwar sediments	Fourth Glacial *Third Interglacial* Third Glacial
	Middle	High terrace Boulder conglomerate	*Second Interglacial* Second Glacial
	Lower	Pinjor zone Tatrot zone	*First Interglacial* First Glacial

Pliocene

correlated with pluvials, and the interglacials with interpluvials. The equivalence of certain Punjab sediments with glaciation is based on their tracing upstream into moraines. The correlation does not seem to fit with Dainelli's implication that only three glacial stages are represented in the Himalayas; reconciliation of the apparent difference awaits further study.

[19] Dainelli 1922, pp. 600, 637–642.

[20] De Terra and Paterson 1939.

A similar sequence, correlated with main divisions in Table 14, has been identified in the terraces and sediments of the Irrawaddy River in Upper Burma.[21]

CHINA, MONGOLIA, AND KOREA

The highland of eastern Tibet, Sikang, and northwestern Yunnan lies at altitudes between 13,000 and 20,000 feet with at least one peak (Minya Konka) exceeding 25,000 feet. Unlike western Tibet, which lies in the rain shadow of the Himalayas, this highland is not shielded from the moist southerly monsoon winds and is therefore not particularly dry. The climate has a high-alpine rather than a steppe character, with heavy precipitation at very high altitudes. As late as 1934 Penck[22] regarded it as an open question whether during the glacial ages this highland had a coalescent glacier cover. Von Wissmann thought coalescent glaciation likely and represented it on a map.[23] On a basis of reconnaissance observations Andersson inferred that the highlands were extensively glaciated.[24] Richardson believed that glaciers had been virtually continuous above an altitude of about 13,000 feet, and that some former glaciers had descended to 10,000 feet. He believed that some of the former glaciers were ice caps.[25]

The Tsinling Shan, an east-west range in central China, south of the Wei River between Lanchow and Sian, reaches an altitude of 10,000 to more than 18,000 feet. The higher groups of peaks are marked by cirques, some of which are at altitudes as low as 14,000 feet.[26]

Evidence of glaciation has been reported from the summit of the Datsin Shan (9400 ft.) northeast of the great bend of the Hwang River in Suiyuan west of Peiping.[27]

In northeastern Korea the Seturei Range, at the northeast end of the Kaima Plateaus, was locally glaciated. The range reaches an extreme altitude of 8382 feet. At altitudes between 6600 and 7000 feet, 18 cirques were formed; their glaciers reached down to 5600 feet. The regional snowline is low here because the mountains receive heavy precipitation from the winter monsoon.[28] Farther southwest in Korea the Shan Alin, with peaks exceeding 8000 feet, likewise was glaciated[29] probably under much the same climatic conditions as the Seturei Range.

[21] De Terra and Movius 1943.
[22] Penck 1934, p. 23.
[23] Von Wissmann 1937.
[24] Andersson 1939, p. 47 *ff.*
[25] Richardson 1943.
[26] Olbricht 1923; Lee 1939, p. 374.
[27] Cited in Antevs 1929*b*, p. 687; the range is there called the Suma Hada.
[28] Sasa and Tanaka 1938.
[29] Berkey and Morris 1927, p. 383.

The Lushan Range, in east-central China between Nanchang and Changsha, has features that have been the subject of much debate. They include such land forms as cirquelike niches, hanging valleys, valleys with U-shaped cross profile, and morainelike topography. They include also deposits interpreted as till and erratic boulders. This assemblage has been taken to indicate glaciation of the range, despite low altitudes which barely exceed 6000 feet.[30] This interpretation, although accepted by Movius,[31] was challenged by Barbour, who thought an early Pleistocene glaciation, with the topographic details obliterated by erosion, a less likely explanation of these features than peculiar structural relations affected by mass-wasting on an extensive scale.[32] The matter must be regarded as still unsettled.

Nonglacial Pleistocene deposits in northern China, including alluvial, lacustrine, and cavern sediments, have been grouped into a general correlation in a useful summary compiled by Movius.[33]

In Outer Mongolia cirques with floors at about 8000 feet occur in the Khangai Range (summits 10,000 ft.) (about 47°00′N, 101°30′E).[34]

AFRICA

The former glaciation of Africa is closely related to the glaciers of the present day, which in turn are dependent chiefly on the distribution of very high land. With one exception, the Atlas Mountains in French Morocco, all the known glaciated highlands lie in eastern Africa within a few degrees of the equator. Three of them lie almost on the equator itself, and three, as stated in Chapter 4, carry glaciers today. Altitudes range from nearly 13,000 feet to nearly 20,000 feet. The snowfall that nourished the glaciers probably was the orographic precipitation of moisture brought by maritime air masses moving eastward from the South Atlantic and, to a lesser extent, moving westward from the Indian Ocean.

The significant data are summarized in Table 15.

Very little information is available on the glaciation of the High Atlas Mountains in French Morocco.[35] These mountains reach a reported extreme altitude of 13,700 feet and carry perennial snow at the present time. The glaciated area lies within the rectangle formed by latitude 30°45′ and 31°25′N, and longitude 7°15′ and 8°50′W.

[30] Cf. Lee 1933; 1938.
[31] Movius 1944, p. 58.
[32] Barbour 1935.
[33] Movius 1944, p. 47. See also Pei 1939.
[34] Berkey and Morris 1927, pp. 130, 384.
[35] Martonne 1924; Roch 1939.

TABLE 15. GLACIATED AREAS IN EASTERN AFRICA [36]

Highland	*Latitude (approx.)*	*Longitude (approx.)*	*Altitude (feet) (approx.)*	*Existing Glaciers*	*Lowest Altitude of Former Glaciation (feet)*
1 Ruwenzori, Uganda	0°24′N	29°54′E	16,912	Yes	7,000
2 Sattimma, Aberdare Range, Kenya	0°19′S	36°38′E	12,870	No	
3. Mt. Elgon, Uganda	1°08′N	34°33′E	14,239	No	<10,000
4. Mt. Kenya, Kenya	0°10′S	37°18′E	17,143	Yes	< 9,800*
5. Kilimanjaro, Tanganyika Terr	3°05′S	37°22′E	19,700	Yes	<10,000
6. Mt Kaka (or Badda), Abyssinia	7°50′N	39°24′E	13,639	No	
7. Semien Mountains, Abyssinia	13°14′N	38°25′E	15,246	No	
8 Mt. Guna, Abyssinia	11°43′N	38°17′E	13,881	No	Probably glaciated
9 Amba Farit, Abyssinia	10°53′N	38°50′E	13,042	No	Probably glaciated
10 Mt. Chillalo, Abyssinia	7°50′N	39°10′E	12,743	No	Probably glaciated

* Glaciated area exceeds 100 square miles.

SOUTH AMERICA[37]

Present-day glaciers in South America are confined to the Andes Cordillera; the former more extensive glaciers had the same general distribution except that in the extreme south, in Patagonia, the ice spread out and coalesced on the plains east of the mountains. The glaciated areas are somewhat irregularly distributed because throughout most of the Andean region they coincide with the distribution of the highest mountains. In general the regional snowline was highest in the regions of latitude lower than 25°. Toward the south it descended with decreasing temperature and in part with increasing precipitation and cloudiness. South of latitude 42° the ice on the west slope of the Andes reached tidewater; that on the east slope reached tidewater only south of latitude 52°. The former glaciers included valley glaciers and small ice caps. South of latitude 26° the ice masses were coalescent, forming a complex that in some regions had the characteristics of a thick mountain ice sheet. In the far south this complex included a great lobate mass that spread out on the Patagonian plains east of the mountains. At its farthest extent, however, the piedmont reached no more than 130 miles

[36] Data from Nilsson 1940; 1935; 1931; P. C. Spink, *unpublished*.

[37] General references are Sievers 1908; 1911; Steinmann 1906. For Columbia see Oppenheim 1940; Notestein *in* Cabot 1939. For Ecuador, Meyer 1907. For Peru, Steinmann 1929, pp. 259–279; Bowman 1916. For Bolivia, Tight 1904; Moon 1939. For Argentina and Chile, Windhausen 1929–1931, vol. 2, pp. 461–478; Walther Penck 1920; Caldenius 1932; Fenton 1921; Quensel 1910.

eastward from the base of the Andes, not much farther than the extent of the Rocky Mountains glaciers eastward over the Great Plains in Canada. The limited eastward extent in both regions was determined by the same factor, the rain shadow of the high mountains.

Deficient precipitation and large summer-season ablation prevented the glaciers from descending far to the west in most latitudes. Only in the far south was spreading of the glacier ice to the west checked by deep water, as in British Columbia and Alaska. Here, on the fiorded Chilean coast, the Andean ice probably formed a narrow floating shelf that calved continually into the Pacific. Glacial and postglacial erosion on the well-nourished western slope of the mountains (where the present-day precipitation reaches 120 inches annually) have been so efficient that almost no drift remains there.

In southern South America nourishment of the glaciers was derived from orographic snowfall from maritime air masses moving eastward from the Pacific. Farther north, however, the winds prevailing at altitudes of more than 10,000 feet are easterly, bringing moisture from the Atlantic and from the Amazon basin. These are believed to bring most of the snowfall received by this part of the Andes. In the southern Andes annual precipitation is comparable with that in the Scandinavian Mountains, but relatively high winter temperatures are a factor in confining snowfall to the higher altitudes.

Evidence of multiple glaciation is very scanty. In the Sierra Nevada de Santa Marta (18,878 ft.) in Colombia only one glacial stage, a very late one, is recognized. It has been suggested that the uplift of these mountains is so recent that they were not high enough to support glaciers during the earlier glacial ages.[38] In Ecuador an old, weathered drift is separated from a fresher drift of smaller extent by a loesslike deposit.[39] Based on degree of decomposition of drifts, two glaciations are recognized in Peru.[40] In southern Argentina four systems of end moraines have been identified, but all are referred to substages of the Fourth Glacial age.[41]

A large amount of loess was deposited east and north of the glaciated region in Argentina, but its stratigraphic relations do not seem to have been generally agreed on.[42] Eolian sediments in that region do not seem to have been clearly discriminated from those deposited by streams and other agencies.

[38] Notestein *in* Cabot 1939, p. 620.
[39] Meyer 1907.
[40] Steinmann 1929, p. 278.
[41] Caldenius 1932.
[42] Cf. summaries in Scheidig 1934, pp. 23, 24; Windhausen 1929–1931, vol. 2, p. 473.

PACIFIC ISLANDS

Aside from the Aleutians, already discussed, and the Australia-New Zealand region, the Pacific islands that were glaciated are Hawaii, Honshu, Hokkaido, Formosa, and New Guinea. In each of these islands glaciation is confined to very high areas and is therefore extremely restricted in extent.

Hawaii

Mauna Kea (13,784 ft.), a volcanic mountain on the island of Hawaii, was glaciated down to about 10,500 feet by a small ice cap believed to date from the Wisconsin stage. The glacial origin of pre-Wisconsin drifts reported on this mountain has been considered doubtful.[43] The present-day snowline is believed to lie at 14,000 to 15,000 feet. During the glacial ages it stood some 2000 feet lower.

Japan

The Japanese Alps[44] lie west of Tokyo in the central part of the island of Honshu. They consist of two masses, the Northern Alps and the Central and Southern Alps. The northern mass, which consists of six distinct ranges, culminates in the peak known as Yari-ga-Také (10,490 ft.). These mountains receive a maximum annual precipitation of 100 inches, most of which is snow. Yet so great is the ablation that none of the snow remains on the ground perennially. Glaciation was correspondingly slight; it involved a large number of small valley glaciers, representing a regional snowline at least 2000 feet lower than the present one.

The Central and Southern Alps, consisting of three ranges, are nearly as high (max. 10,382 ft.) as their neighbors but are less favorably situated to receive heavy snowfall. Probably this fact explains their lesser glaciation, which involved fewer than a dozen small cirque glaciers. East of the Southern Alps Fujiyama (12,467 ft.), the highest peak in Japan, shows no evidence of glaciation. As with other volcanic cones of recent date, the explanation may lie in the destructive effects of postglacial volcanic activity.

In the central part of the island of Hokkaido the Hidaka Mountains reach an extreme altitude of 6756 feet. This highland contains thirty cirques, with floors at 4800 to 5300 feet. Most of the cirques are on the northeast flank of the mountains. The comparatively low level of glaciation in this range, as compared with the Japanese Alps, results from the fact that it lies 8 degrees of latitude farther to the north.[45]

[43] See H. T. Stearns 1945*a* and references therein cited.

[44] Imamura 1937. A less detailed but more generally available reference is Oseki 1915.

[45] Sasa 1934.

Formosa

In southeast-central Formosa (lat. 24°10′ to 24°30′ N; long. 121°00′ to 121°20′E) there is a northeast-southwest range of mountains more than 25 miles long. The highest peak is Mt. Tugitaka (also known as Mt. Morrison and as Mt. Sylvia) with an altitude of 12,972 feet. The summit of the range is a narrow plateau, indented by a total of thirty-five cirques along its northern and eastern margins. The cirque floors lie at altitudes of 11,500 to 12,300 feet, and other glacial features extend down to 10,900 feet. As in the Northern Alps on Honshu, precipitation now reaches 100 inches annually, but there is no perennial snow. Only the altitude of this highland made glaciers possible. The glaciation is correlated with the Fourth Glacial stage.[46]

New Guinea[47]

Though very close to the equator, and though clothed throughout much of its extent with tropical rain forest, the island of New Guinea nonetheless possesses two highlands that have been glaciated; one of them supports a small ice cap today.

The lower and lesser of the two highlands is the Saruwaged Range (6°20′ S; 147°05′ E) whose culminating peaks are Mt. Bangeta (13,470 ft) and Mt. Salawaket (13,454 ft.). This range formerly supported small glaciers, as recorded by cirques. The greater highland lies farther west (4°05′ to 4°15′ S; 137°10′ to 138°30′ E) and includes the Nassau Range and the Emma Mountains. The highest peak, Ngga Poeloe, reaches 16,600 feet. The existing glaciers on these summits were formerly much larger, one valley glacier having been 10 miles long, descending to 6600 feet above sealevel. According to Dozy the atmosphere is so moist that this expansion could be brought about by a very small reduction in the present mean annual temperature.

AUSTRALIA

Most of the Australian highlands are too low to have reached the regional snowline even during the glacial ages when it was at its lowest. An exception is the Kosciusko Plateau, south of Canberra in southeastern New South Wales. Much of this mountainous highland exceeds an altitude of 6500 feet, and the culminating point, Mt. Kosciusko, reaches 7308 feet. Annual precipitation is about 60 inches, derived principally from air masses approaching from the west.

[46] Kano 1934–1935; 1940.
[47] Behrmann 1928; Dozy 1938.

There is abundant evidence of a former ice cap with a maximum area of more than 150 square miles and descending to 5000 feet. When the ice cap existed the regional snowline lay at least 3000 feet lower than it does today. At a later time small valley glaciers formed, as recorded by cirques with floors as low as 5800 feet. The earlier, ice-cap, glaciation has been tentatively correlated with the Third Glacial stage; the later, cirque, glaciation with the Fourth.[48] It seems possible, however, that the earlier glaciation may date from the earliest sub-age of the Fourth Glacial age.

In Victoria, some 50 miles southwest of the Kosciusko Plateau, Mt. Bogong (6509 ft.) is said to show evidence of local glaciation.[49]

As far as is known, this is the extent of glaciation in Australia (Fig. 65). Widespread glaciation, however, occurred in Tasmania. The western part of this island is a dissected plateau, with an extreme altitude of 5069 feet in Cradle Mountain. The annual precipitation near the west coast reaches 100 inches and is derived from moist maritime northwesterly winds. Tasmania has no glaciers today, but at the time of maximum glaciation a complex of ice caps and valley glaciers centered on this highland and covered at least 10,000 square miles and attained a thickness of more than 1500 feet. Outlet glaciers reached the sea at various points along the west coast. At that time the regional snowline may have been at least 4000 feet lower than its present altitude of 6000 feet or more (Fig. 65).

Even though slightly higher altitudes are found along the east coast, glaciation in that region, which lies in the rain shadow of the western highlands, was confined to a single summit, Ben Lomond (5160 ft.).[50] The shadowing effect would have been greatly accentuated during the glacial ages.

Interglacial climates are suggested by peat beds occurring between drifts, and a long time between successive glaciations is recorded by the profound erosion of bedrock valleys by streams between episodes of drift accumulation. The erosional relations are like those between the Kansan drift and the later drifts in the northern part of the Rocky Mountains in western United States.

The extensive glaciation was followed by two separate episodes of valley glaciers. The glacial history is summarized in Table 16. The three glacial units shown in the table were regarded as stages and were tentatively correlated with the Second, Third, and Fourth glacial stages of Europe. However, it seems more probable that the Malanna represents

[48] Browne and Dulhunty 1944; Browne 1945 (containing a correlation chart showing Pleistocene events in Australia).

[49] David 1932, p. 94.

[50] Nye and Lewis 1927, p. 25.

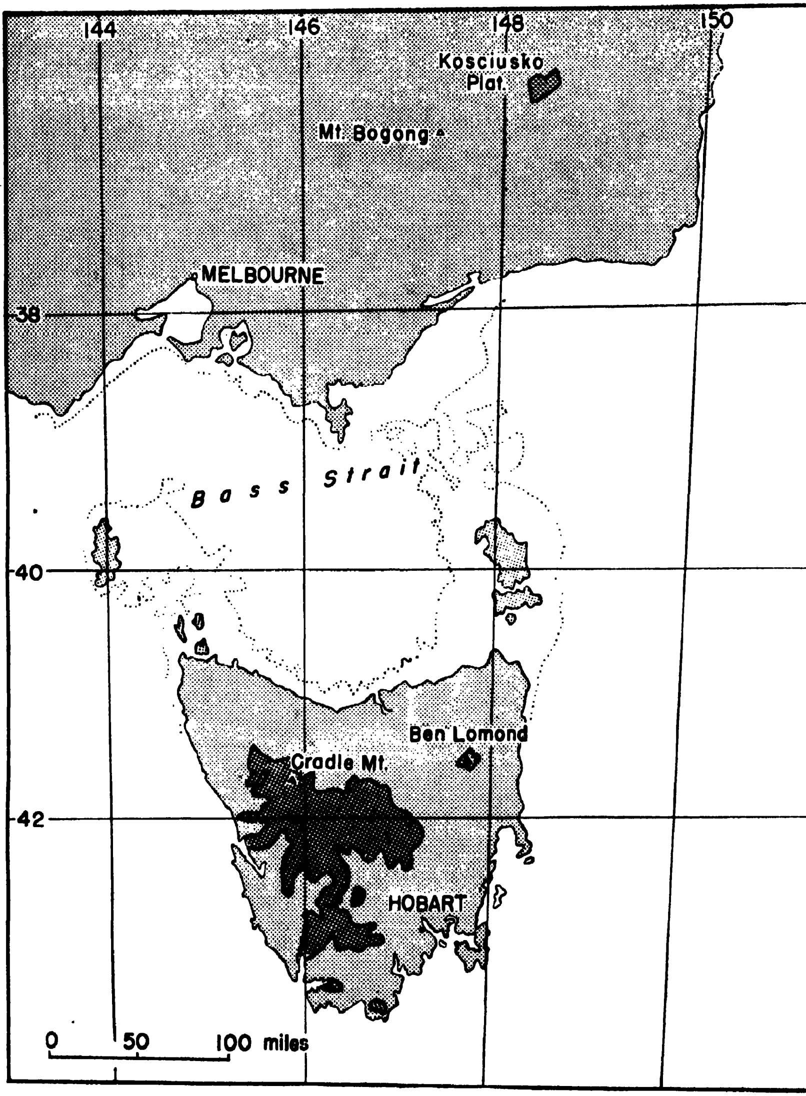


FIG. 65. Glaciated areas in Australia and Tasmania (dark shading). Dotted line = form line at 180 feet below sealevel. (Modified from maps by W. N. Benson and W. R. Browne.)

TABLE 16. PLEISTOCENE SEQUENCE IN TASMANIA ACCORDING TO LEWIS[51]

Unit	*Event*
Margaret	Small valley glaciers constituting a new generation.
Interglacial	Minor erosion.
Yolande	Valley glaciers with distinct end moraines.
Interglacial	Extensive dissection of Malanna drift; 2000 ft. of valley cutting in resistant bedrock.
Malanna	Extensive drift related to ice-sheet complex. Now greatly eroded.
	(Pre-Malanna events are obscure)

the Third Glacial stage and that the Yolande and Margaret are substages of the Fourth Glacial stage. The Malanna stage, further, would seem to correlate with the ice-cap glaciation of the Kosciusko Plateau in New South Wales.

NEW ZEALAND[52]

On the North Island of New Zealand only minor glaciation occurred (Fig. 66). The volcanic cone Ruapehu (9175 ft.) (39°20′ S; 175°40′ E),

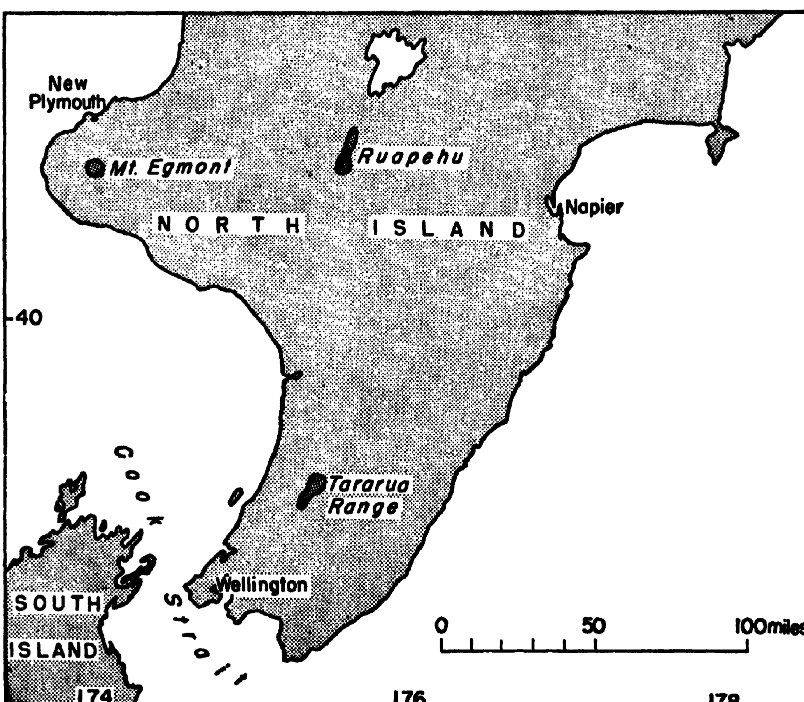


FIG. 66. Glaciated areas (dark shading) on the North Island, New Zealand. (Data from Willett 1940.)

[51] A. N. Lewis 1934.
[52] Speight 1939; Willett 1940.

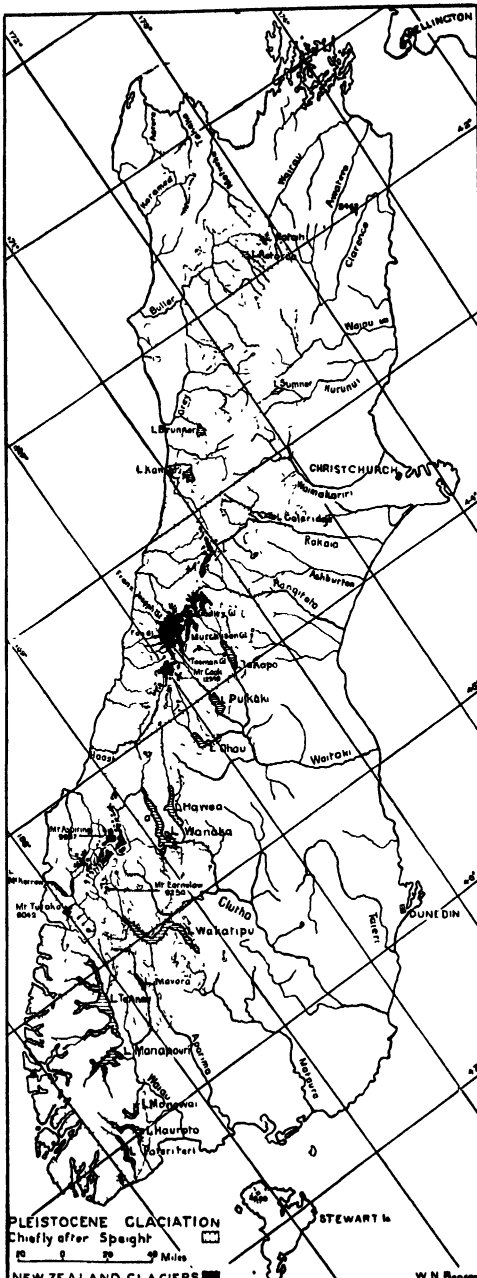


FIG. 67. Existing glaciers and areas of former glaciation on the South Island, New Zealand. (W. N. Benson.)

whose crater is today filled with glacier ice, was more extensively glaciated, with the development of cirques having floors at 8000 feet.[53] Mount Egmont (8260 ft.), another volcano on the west coast at latitude 39°20′ S, longitude 174°00′ E, nourished small glaciers. Farther south the Tararua Range (5154 ft.), in the southern part of the island, exhibits evidence of having been mildly glaciated.

It is in the mountainous western part of the South Island, the New Zealand Alps, that glaciation was both widespread and intense (Fig. 67). Sizable glaciers exist today in the higher parts of these mountains which culminate, in Mt. Cook, at an altitude of 12,349 feet. Formerly glaciers coalesced into a vast complex of mountain ice sheets and valley glaciers, the whole mass more than 400 miles long and with an average width approaching 50 miles. Great piedmonts encroached over the plains east of the mountains. To the west the ice reached the fiord-cut coast as thick valley and piedmont glaciers and very probably coalesced to form a stout though narrow floating shelf offshore. Glacial erosion was intense, reflecting the abundant snowfall that resulted from the ideally maritime position of the New Zealand Alps.

Stewart Island, south of the South Island, had only a small area of glaciation on its highest part, 3200 feet above sealevel.

In several parts of the South Island two glacial stages have been identified on evidence furnished by moraines and outwash. No interglacial deposits containing fossil plants or animals have as yet been found. The intense glaciation of the west coast explains this in part, though the less deeply eroded east flank of the mountains, and the plains east of them, may in the future yield important evidence of this kind.

ANTARCTIC AND SUBANTARCTIC ISLANDS

Table 17 summarizes the glacial information about the larger islands in Antarctic and subantarctic waters. From it appears the fact that virtually all these lands now have glaciers; all or nearly all of them were formerly more extensively glaciated, some of them completely so.

Macquarie Island[54]

Macquarie Island, south of New Zealand, requires special mention because it presents a peculiar problem. Surrounded by deep water, this island reaches an extreme altitude of 1421 feet. It has been completely glaciated. It is apparent, however, from the distribution of striations and

[53] Griffith Taylor 1926, p. 1821.
[54] Mawson 1943, pp. 45–54, 76–83, 92.

indicator stones that the former ice did not spread out from a center on the island but crossed the island from a center farther west where now there is only deep water. Not only were stones carried to the east coast from bedrock outcrops on the west coast; they were lifted 1200 feet vertically in the process. The glacier must therefore have been large. The present island is regarded as a residual horst left by profound postglacial block-faulting movement in a land mass that has largely disappeared.

TABLE 17. CHIEF SUBANTARCTIC AND ANTARCTIC ISLANDS: GLACIAL DATA [55]

Name	S. Latitude (approx)	Longitude (approx)	Area (sq mi.)	Extreme Altitude (feet)	Existing Glaciers	Former Glaciation
Auckland Is.	51°30′	166°00′E		2000+	No	Partially glaciated
Macquarie I.	54°30′	158°57′E		1421	No	Entirely glaciated
Heard Is.	53°30′	73°30′E	135	6000	Ice cap	Yes
Kerguelen Is.	49°25′	69°53′E	1400	6500	Yes	Extensive ice cap
Crozet Is.	50°00′	46°30′E		5000+	Yes	Yes
Bouvet I.	54°30′	3°30′E		3000	Yes	Present glaciers cover the island
South Georgia	54°-55°	36°-38°W	1600	7300	Yes	More extensive than now
South Sandwich Is.	55°-60°	25°-28°W		4500	Yes	(Actively volcanic)
South Orkney Is.	61° 00′	45°-46°W	155	3000	Yes	Extensive
South Shetland Is.	62°-63°	58°-61°W		6800	Yes	

ANTARCTIC CONTINENT

As was stated in Chapter 4 the Antarctic Ice Sheet was formerly at least a thousand feet thicker than it is now and all the islands fringing Antarctica were extensively glaciated. The former greater glaciation of the peripheral lands certainly coincided with the glacial ages, and it seems likely that the glaciation of the mountains in the interior by the much-thickened central body of the ice sheet took place at the same times. However, opinions on this point differ, as has been set forth elsewhere, and the matter is undecided.

Most of the very small area of Antarctica not now covered by glacier ice is bare glacially eroded bedrock largely free of drift. But submarine ridges, such as the Pennell and Iselin banks extending across the mouth of the Ross Sea, have been held to be a series of submerged end moraines.[56] Some of them stand so high that bergs run aground in the shoal water above them.

[55] The best general reference is *Discovery reports,* published in many volumes by Cambridge University Press.

[56] Mawson 1935, p. 30.

Core samples from the bottom off the Antarctic coast have been identified as marine till. This material mantles the sea floor throughout a zone 200 to 500 miles wide surrounding the Antarctic Continent. Probably much of it was deposited from shelf ice and bergs.[57]

INTERCONTINENTAL CORRELATION

For obvious reasons the correlation of drift sheets by means of continuous tracing can not be carried beyond the limits of one continent. Therefore, in order to establish correlations between continents, especially between Eurasia, North America, and South America, other means must be found. At least two methods are possible.

One method is the comparison of drifts based on stratigraphic similarities, especially the extent to which they have been decomposed. As set forth in Chapter 16, attempts to do this have been made,[58] and further attempts should produce good results.

A second method is the study of closely spaced core samples of sediments taken from the floors of the oceans between continental areas in which the sequence of glacial and interglacial deposits is well known. The deep-sea record has the great advantage of being a continuous one, free of all the gaps that plague the pursuit of glacial-stratigraphic studies on the lands. Furthermore, sea-floor sedimentation should be at least as sensitive to climatic changes as would the regimen of a large ice sheet. The results should therefore have a value second only to that of the impossible ideal of continuous tracing. The preliminary results already obtained by this method are described in Chapter 20.

The application of these two methods of research should result in the establishment of correlations between glacial stages throughout the glaciated regions of the world. The prospects of the correlation of substages are less good because of the possibility, already mentioned, that substages even in different sectors of the margin of a single ice sheet may not have been contemporaneous. With more detailed study, however, this difficulty may disappear.

[57] Stetson and Upson 1937.
[58] Cf. Leverett 1910; see also Nilsson 1941.

Chapter 18

PLEISTOCENE CHRONOLOGY

One of the great geologic advances of the twentieth century has been the development of a method of dating more or less accurately, in terms of actual years, definitely known points in geologic history. As a result our perspective of geologic time has been vastly enlarged, and we have gained a more accurate concept of the time involved in the evolution of animals and plants. Within the last few decades attempts have been made to arrive at an absolute time scale for Pleistocene events, with all that the possession of such a scale would imply for an understanding of the growth and decay of vast glaciers and of the evolution of man. Although a good deal of ingenuity has been expended on the matter, the results thus far obtained are only estimates, and all are unsatisfactory because they are unavoidably inaccurate. The best of them have hardly yielded more than an idea of the order of magnitude of the time involved.

Ever since Agassiz proposed the glacial theory geologists have inferred that the glacial ages came very late in the geologic record, because glacially smoothed and striated bosses of bedrock were seen preserved in virtually original freshness. Soon afterward morainic topography, unfilled small basins that had contained lakes, and similar evidences of geologic recency were noted, but all such features are far from giving any precise concept of the length of time since they were made, because the factors involved in their preservation are variable. In Yosemite Valley, in places where silicious granite is in contact with diorite, the granite "gleams with glacier polish, whereas the diorite has a roughened and perceptibly lower surface."[1] If these two rock types were not in contact, but occurred in different districts within the same region, their different surfaces might have been interpreted as the products of two different glaciations. In this instance the variable is lithology. In another it may be climate; in still a third, angle of slope. But always, even with the greatest caution, the results of our calculations are necessarily coarse.

The attempts at an absolute rather than a purely relative chronology of Pleistocene events fall into two groups: those aimed at the determination of postglacial time, and those looking toward the dating of events

[1] Matthes 1930, p. 69.

farther back in the Pleistocene. The majority are in the first group, seeking to determine the time that has elapsed since some identifiable event in the last deglaciation. The first group is the larger mainly because the available evidence is clearer than for earlier events.

POSTGLACIAL CHRONOLOGY

Postglacial is a term that, strictly speaking, can be applied to only one locality or district. Events that took place while the district was last covered with glacier ice are referable to glacial time; subsequent events are referable to postglacial time. Since removal of the glacial cover was progressive, it follows that postglacial time began for some districts much earlier than for others, and that for most of Antarctica and Greenland, for example, postglacial time has not begun yet. The term postglacial is sometimes applied to events in regions that never were glaciated but in which the indirect effects of glaciation are clearly marked, as for example the valleys of large streams whose headwaters drain glaciated regions. The cessation of outwash deposition and the beginning of trenching of the outwash by nonglacial runoff is an event that can often be definitely recognized and that can with propriety be considered as marking the transition from glacial to postglacial time in a given valley or valley system. Where there is no direct connection, such as is afforded by outwash, between a glaciated and a nonglaciated district, the term postglacial is hardly applicable to any feature in the nonglaciated district. For the desert basins of western United States, where Pleistocene features record alternating episodes of moisture and dryness, terms such as *pluvial* and *postpluvial* are sometimes used.[2]

The length of postglacial time in any given region has been approached in two ways: through the study of varved lacustrine sediments and through extrapolation backward on measured or estimated rates of natural processes still in progress. We shall begin with the second.

Extrapolation on Rates of Present-Day Processes

All attempts at extrapolation either carry the expressed or implied assumption that the rate has been constant or must make allowances for past variations in rate. Further, the time of inception of the process usually can not be definitely fixed. These disadvantages are common to the group; in addition, specific problems involve other, individual difficulties.

[2] Cf. Antevs 1936.

Rate of Erosion

Retreat of waterfalls: Niagara. Niagara River cuts through a cuesta that separates the basin of Lake Erie from that of Lake Ontario. The cuesta is formed by a resistant dolomite overlying a weak shale. Largely by undermining of the dolomite, the falls have migrated 6.5 miles up the

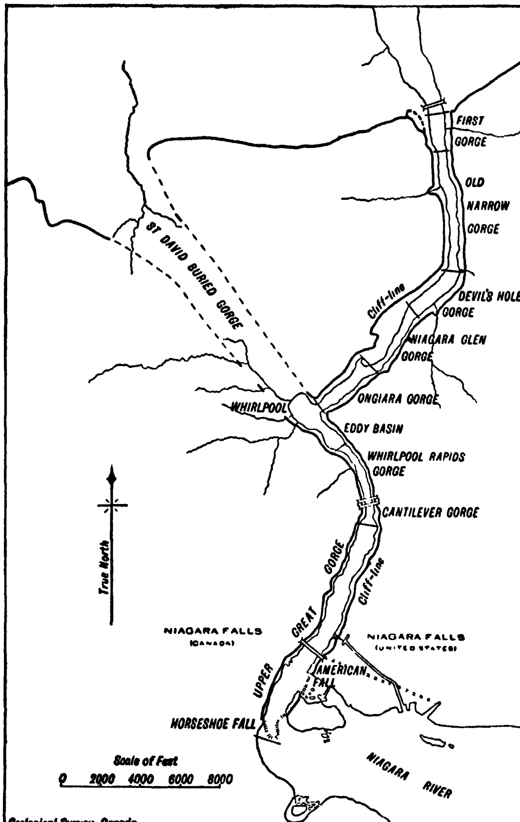


Fig. 68. Plan of Niagara Gorge and surroundings (F. B. Taylor, U. S. Geol. Survey).

river from the cuesta scarp, leaving a gorge of variable width through this distance (Fig. 68).

The falls are believed to have come into existence at a time when an ice sheet, shrinking away toward the north, uncovered the cuesta and permitted water that had been ponded by the ice in the Lake Erie basin to spill northward into the much lower basin of Lake Ontario newly freed of glacier ice. At first it was believed that these events occurred entirely during the Wisconsin age. On this basis Gilbert wrote: "To

measure the age of the river is to determine the antiquity of the close of the ice age."[3] For a long time it was thought that the chief fall, the Horseshoe Fall, was retreating at a rate of about 5 feet per year, and that dividing this annual rate into the total length of the gorge would measure postglacial time for the Niagara district. Then it was found that the discharge of Niagara River had amounted at times to more than twice its present discharge and at others to a small fraction of this quantity. Further it was realized that the level of the lake in the Ontario basin had been high enough to reduce the drop of the falls by half. Both these factors introduced important variables into the calculation, and, despite ingenious estimates of the discharge at various times, the closest time estimate of the age of the gorge that could be made in 1913—20,000 to 35,000 years[4]—involved a large margin of error.

It was shown later, however, through borings made for a railroad bridge, that the middle part of the gorge, the Whirlpool Rapids Gorge (Fig. 68), contains a thick fill of glacial drift, and hence that it must be a pre-Wisconsin (probably interglacial) gorge that was filled with drift and then partly re-excavated in postglacial time.[5] If this is true, then the age of Niagara Gorge as a whole can not be determined, as we have no way of fixing either the rates of cutting or the dates of the parts of the gorge that antedate the last glaciation. We are obliged to fall back on the Upper Great Gorge, the uppermost segment of the whole gorge, which appears to be genuinely postglacial. Redeterminations by W. H. Boyd showed the present rate of recession of the Horseshoe Fall to be not 5 feet, but rather 3.8 feet, per year.[6] Hence the age of the Upper Great Gorge is calculated at somewhat more than 4000 years—and to obtain even this figure we have to assume that the rate of recession has been constant, although we know that discharge has in fact varied greatly during postglacial time. Accepting Johnston's data, Taylor revised his interpretation;[7] this, however, added nothing in the form of exact chronology. In summary, the history of Niagara Falls as we now understand it does not furnish a basis for determining the duration of postglacial time.

St. Anthony Falls. Like Niagara, the St. Anthony Falls of the Mississippi River at Minneapolis are related to postglacial drainage; like Niagara, the gorge below the falls is nearly 7 miles long; and like Niagara, the downstream end of the gorge is an exhumed valley of pre-

[3] Gilbert 1895, p. 234.
[4] F. B. Taylor 1913, p. 24.
[5] Johnston 1928
[6] Johnston 1928.
[7] F. B. Taylor 1929.

Wisconsin date. Like Niagara, these falls have been made the basis of estimates of postglacial time.

In pre-Wisconsin time the Minneapolis-St. Paul district in Minnesota had a system of stream valleys (Fig. 69) with which the modern valleys are coincident only in part. Wisconsin glaciation completely filled the old valleys with drift, and the meltwater streams of earliest postglacial

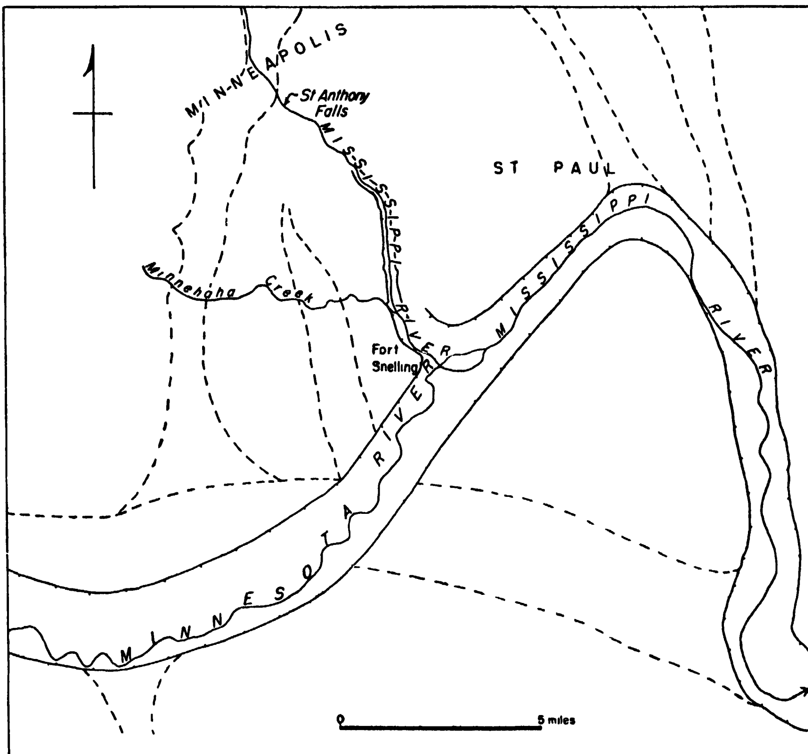


Fig. 69. Plan of vicinity of Minneapolis-St. Paul, showing St. Anthony Falls and features referred to in the text. (Modified after G. M. Schwartz, Minnesota Geol. Survey.)

time took new routes over the drift surface. In time the postglacial main stream discovered the buried valley in the southeast part of St. Paul, rapidly re-excavated it, and, where it flowed over the side of the exhumed valley, formed a falls. As this falls retreated past Fort Snelling, where the Mississippi River joins the Minnesota River, it cleared out a narrow preglacial valley there, and a falls was started in the Mississippi about a mile north of Fort Snelling. This falls, St. Anthony Falls, retreated northward through nearly 6 miles, and still exists, but the main falls,

retreating southwestward, intersected another preglacial valley and was thereby extinguished.

Through nearly 3 miles above Fort Snelling the Mississippi Gorge is deep and contains a thick fill of alluvium. The 4-mile segment extending thence upstream to the St. Anthony Falls is shallower and contains no alluvium. The difference arises from the fact that the Minnesota River valley first carried the enormous flow of water discharged from Lake Agassiz, and later, when Lake Agassiz drained northward, this discharge ceased, and aggradation by the reduced Minnesota River was the result. The Mississippi, which up to that time had been a mere tributary to the swollen Minnesota, aggraded its valley floor to the rising fill in the master valley.

By comparing maps, sketches, and descriptions dating back as far as 1680, it has been computed that the St. Anthony Falls has retreated at an average rate of 2.44 feet per year. On this basis Winchell calculated that about 8000 years had elapsed since the falls originated.[8] Sardeson later calculated this period as 12,000 years, of which 8000 years have elapsed since the draining of Lake Agassiz and the consequent aggradation of the Minneapolis-St. Paul district.[9]

Unfortunately for the accuracy of these calculations, the rate of recession of the falls was found to have accelerated notably during the time of human observation. Probably this is because the Platteville limestone, the resistant stratum that forms the lip of the falls, thins rapidly in the upstream direction. This factor and the factors of variable discharge and falls height introduce variables into the calculation that can not be overcome. Hence the figures given in the preceding paragraph are not only not accurate but may not even be of the right order of magnitude.

Summarizing the studies of the recession of Niagara and St. Anthony falls, we are obliged to conclude that neither measures the whole of postglacial time for the district in which it occurs, and that for both the results are unreliable because of variable factors that can not be evaluated.

Retreat of wave-cut cliffs. Somewhat analogously, Coleman made an attempt to estimate the age of Lake Ontario, which, as he recognized, represents only a part of post-Mankato time. He inferred that off the Scarboro cliffs, near Toronto, the shore initially stood 13,000 feet farther out than now, because at that distance the offshore bottom profile abruptly steepens. By comparing the results of successive surveys made during a period of 50 years, he calculated that the cliffs had receded at an annual rate of 1.62 feet. At this rate the inferred initial shoreline of Lake Ontario would date back to about 8000 years ago, a figure which Cole-

[8] Winchell 1888, pp. 337, 341.
[9] Sardeson 1916, p. 13.

man regarded as the minimum. Accretions of debris from the erosion of the cliffs to the shore farther west were calculated to have required a comparable time for their development.[10]

The Lake Iroquois strandline is as maturely developed as that of the modern Lake Ontario; by analogy Coleman inferred that its development had likewise required about 8000 years. Several lake phases intervened between Iroquois and Ontario, and, in the absence of any time data on these, Coleman guessed that another 8000 years should be assigned to them, making a grand-total estimate of 24,000 years elapsed since Lake Iroquois time, a time when glacier ice still covered most of southeastern Canada.

This estimate was criticized by Spencer.[11] Although we may think his criticism unnecessarily severe, we are obliged to agree that indeed two-thirds of Coleman's result is guesswork, whereas the other third involves several uncertainties. All in all, it is hardly as trustworthy even as the unsatisfactory results of the studies of falls recession.

Erosion of small valleys. G. F. Wright attempted to extrapolate backward on the rate of cutting of an artificially diverted creek in Ohio during a period of observation of 14 years, in order to determine the period required for the excavation of the stream's natural valley, a period just equal, in Wright's opinion, to postglacial time for the district in question. Wright explicitly recognized two variable factors: (1) the decrease in rate of erosion with increasing topographic age of the valley, and (2) the strong influence of forest cover, which was present during most of the cutting of the natural valley but not during the cutting of the valley begun by the diversion. In consequence, the figure arrived at, 2500 years, was admittedly much too small and is of little value.[12]

In summary, extrapolation backward on present rates of erosion, even where those rates are fairly closely fixed, is an unsatisfactory method for determining the length of postglacial time. We may now turn to a series of attempts based on rates of sedimentation.

Rate of Sedimentation

Delta building. One of the sedimentary processes upon which time estimates have been constructed is the building of deltas. Johnston studied the delta of the Fraser River, which enters the Pacific at Vancouver, British Columbia, with the object of estimating the time involved in its accumulation.[13] The modern delta of the Fraser lies within and

[10] Coleman 1914.
[11] Spencer 1917.
[12] G. F. Wright 1912.
[13] Johnston 1921.

below the remnants of a dissected delta built during a glacial time when the land stood at least 650 feet lower, relative to the sea, than it does now. The modern delta was built after uplift of the land relative to the sea had ceased. It is 80,000 feet long, and in one place it is at least 700 feet thick. Measurements made through a 60-year period show that the average rate of seaward advance of the delta front is about 10 feet per year. On the assumption that this rate has been uniform throughout the existence of the modern delta, the age of the delta is 8000 years. The chief variables that have to be neglected in this calculation are variable rate of sedimentation and variable width and depth of the submerged valley in which the delta is built. Both variables could introduce large errors into the result. Furthermore it seems very likely that, after relative uplift of the land had ceased, the sealevel continued to rise eustatically, perhaps for a long time. As long as it continued, such rise would add to the thickness of the delta and reduce the rate of its seaward advance, thus lengthening by a factor that can not now be determined the elapsed-time figure derived solely from the present rate of growth. Because of these variables, the figure of 8000 years can hardly be regarded as even approximate, and even then it is not certain that the delta began to be built as soon as the Vancouver district was deglaciated.

Applying a similar method on the Bear River delta at Stewart, in extreme northern British Columbia, Hanson[14] derived a figure of only 3600 years. The significance of this figure is discussed in Chapter 21.

The same method was used earlier by Heim on the delta built by the river Muota into Lake Luzern in Switzerland. He calculated that this delta, whose building began with the deglaciation of the locality during the Bühl sub-age, had been accumulating for 16,000 years.[15] The variables applicable to the Fraser delta apply here as well. Using somewhat different values for certain factors, Collet derived a figure of 13,000 years, which he considered more accurate than Heim's figure.[16]

From an examination of these examples we find that in the study of deltas, as in the study of rates of erosion, it is thus far impossible to eliminate important variable factors; the results achieved must be viewed in the light of this fact.

Accumulation of peat. A 39-foot section of sphagnum-moss peat containing spruce stumps, resting on fresh till, at a point 8 miles beyond the terminus of Russell Glacier in Alaska provided Capps with a means of estimating postglacial time at that locality.[17] The peat was frozen

[14] Hanson 1934.
[15] Heim 1894.
[16] Collet 1925, p. 256.
[17] Capps 1916, p. 69.

to within 6 inches of the surface. The roots of the spruces growing on the surface of the peat and the roots of the spruce stumps occurring throughout the peat section were alike in consisting of a central vertical root with several radial horizontal branches. Evidently each radial set of root branches had been sent out just above the frost line, but the thickening of the moss and the accompanying rise of the frost line froze these roots and forced the tree to send out a higher set (Fig. 70). If this is what happened, then the vertical distance between the lowest roots of a living tree and the surface of the ground represents the thickness of peat accumulation during that tree's life. The ages of the living trees being determined by counts of their annual rings, it was calculated that peat is accumulating at a rate of 1 foot in 200 years, the

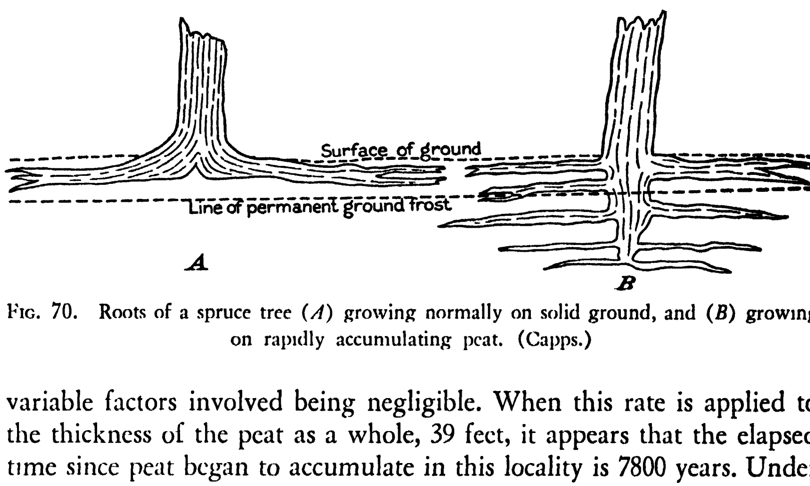


FIG. 70. Roots of a spruce tree (*A*) growing normally on solid ground, and (*B*) growing on rapidly accumulating peat. (Capps.)

variable factors involved being negligible. When this rate is applied to the thickness of the peat as a whole, 39 feet, it appears that the elapsed time since peat began to accumulate in this locality is 7800 years. Under prevailing climatic conditions this would be equivalent to the time since deglaciation of the place. As the rate of peat accumulaton is not likely to have been as variable as that of delta sediments, this figure is probably more nearly of the correct order of magnitude than figures based on deltas.

SEDIMENTATION ON LAKE FLOORS. Many years ago a plan was made[18] to determine the age of Lake Mendota, Wisconsin, which has been exposed since the Cary sub-age, but the project was never carried out. Calcareous mud is accumulating on the lake floor, and the plan was to determine the rate of sedimentation by means of pans placed in various parts of the lake, and the thickness of accumulated sediment by core borings. The physical difficulties of sampling plus the variables that could not be evaluated in the computation make this a hazardous way to

[18] Hotchkiss 1917.

try to determine the length of postglacial time. Perhaps it was because of this that the plan was never implemented.

PRECIPITATION OF TRAVERTINE. Swinnerton made a preliminary estimate of the time since the deglaciation of Yellow Springs, Greene County, Ohio, in the Tazewell sub-age, based on the precipitation of travertine at the springs.[19] Assuming that glacial erosion had removed all preexisting travertine, he divided the volume of existing travertine by the observed annual rate of accumulation and arrived (tentatively) at 20,000 to 30,000 years.[20] He recognized variations in discharge, solution, rainfall, temperature, and subsurface conduits as important variables. This method may be applicable to other spring deposits. Unfortunately Swinnerton did not publish the details of his computation.

Rate of Radioactive Disintegration of Travertine

Before modern studies of radioactivity as applied to geologic time began, Schlundt and Moore, as an incident in the study of the radioactivity of thermal waters in Yellowstone National Park, attempted to date the radium-bearing travertine of Terrace Mountain.[21] This travertine is one of the earliest deposits at Mammoth Hot Springs and is overlain by boulders deposited by the latest glacier ice at this locality. They found that this early travertine contains only about 1 per cent of the radium contained in the travertine now being deposited. Assuming that both deposits originally contained the same amount of radium and that none of the radium content in the earlier deposit had been lost by leaching, and using the best value for the half-time period of radium that was available in 1909, they derived a figure for the age of the early travertine. Using the more recent figure of 1690 years for the half-time period of radium, we may follow their method and derive an age of 11,200 years for the early travertine.

This general method is based on the unproved assumption that the rate of deposition of travertine has been constant throughout the time involved. In addition to this disadvantage, occurrences of radium-bearing deposits of Pleistocene date are very rare. Nor can Pleistocene igneous rocks be made use of, because the lead and helium methods of age determination, applicable to rocks of this type, are not applicable to the immediate geologic past. Pleistocene time is so short compared with earlier recognized units of geologic time that it falls within the limits of error of these methods of determination.

[19] Swinnerton 1925.

[20] The time since the Tazewell was estimated by Thornbury (1940) at 45,000 years.

[21] Schlundt and Moore 1909, p. 33.

Summarizing the results obtained by extrapolation on the rates of various processes, we find that none of them is very satisfactory because all include variable factors that can not be evaluated, and further because, in most cases, the time of inception of the process concerned can not be definitely fixed.

Calculation from Varve Counts

The use of varve counts in the attempt to arrive at the length of post-glacial time in any region avoids the necessity of extrapolation because it does not depend on present rates. Instead it utilizes deposits that were laid down during much of the time whose length it is desired to measure.

Origin of Varved Sediments

The glaciated regions in all continents include fine sediments, chiefly clay and silt, laminated in a peculiar way. Ideally the laminae occur in pairs, each pair consisting of a coarse-grained member (chiefly silt) and an overlying fine-grained member (chiefly clay). The coarse member grades up into the fine member but is sharply separated from the fine member of the next underlying pair. Thus it is possible, even in a hand specimen, to determine top and bottom in the layers. The colors of the chief minerals that constitute the silt—quartz and feldspars—are pale; hence the coarse or silt members stand out in contrast with the darker-colored fine or clay members. A clean exposure thus has a striking uniformly banded appearance (Fig. 71).

The paired arrangement of the layers early led to the speculation that this character is the result of periodic changes in the conditions of deposition, such as occur with the year's changing seasons.[22] Hitchcock, Emerson, Upham, and Taylor all recognized this relationship, but the detailed study of such deposits has been the work chiefly of Gerard De Geer, beginning in 1882. In 1912 De Geer proposed calling each pair of layers, which he recognized as annual, a *varve* (Swed. *varv* = a periodic repetition or revolution).[23] Since that time deposits having this paired characteristic have been rather generally called *varved clay,* an inappropriate term in that usually more than half the deposit consists of silt and other particles coarser than clay.

Johnston showed that the varve period in the sediments now being deposited in Lake Louise, Alberta, is the year rather than any larger or smaller division of time,[24] and agreement is now general that most of

[22] Cf. Alfred Smith 1832, p. 229.
[23] De Geer 1912, p. 242.
[24] Johnston 1922, p. 383.

R. F. Flint, Connecticut Geol. Survey

FIG. 71. Exposure of varved silt and clay, South Hadley, Massachusetts.

the varves apparent in glacial sediments are in fact annual. The geologic evidence shows that these deposits were laid down on the floors of lakes fed by meltwater from the waning glaciers, mainly though not entirely during the last deglaciation.

The conditions of deposition have been inferred from the geologic evidence and also from the results of laboratory experiments, as, for example, the excellent studies made by Fraser.[25] It is clear that suspended sediments, consisting chiefly of silt and clay, were brought into a temporary glacial lake by streams both from the melting ice and from the surrounding land during the spring and summer seasons. The silt, being heaviest, settled out fairly soon, but the finer clay particles remained in suspension, settling out slowly and gradually during the autumn and winter, when melting had ceased. This gradual separation caused the observed gradation from the coarse member to the fine member of the varve pair. The low temperature of the glacial water was an important factor in delaying the settling of the clay particles. With the resumption of melting in the spring, new sediment began to be poured into the lake, and the coarse fraction of it settled out rapidly on the lake floor, producing the relatively sharp contact between the top

[25] Fraser 1929.

of the older varve pair and the base of the younger pair. Thus the varve year commences with the spring.

The salts in seawater prevent the formation of varves of this kind in marine sediments. Being electrolytes, they cause flocculation of the sediment suspended in the water. The sediment is then precipitated as a homogeneous mass of mixed coarse and fine particles, and varves can not form.

The prime condition necessary for varve formation, then, is that much of the fine sediment must remain in suspension throughout the spring and summer seasons. Therefore the water must be fresh and cold, and the stream water entering the lake must be lighter than the bulk of the lake water so that it can diffuse over the whole area of the lake. These requirements explain why varves are far better developed in glacial-lake sediments than in lake sediments of nonglacial origin. Undoubtedly, then, the vast majority of the known bodies of varved sediment are the product of glacial meltwater deposits and hence record the near presence of glacier ice where and when they were laid down.

Measurement and Correlation

Acting upon the very probable assumption that the glacial-lake varves exposed in southern and central Sweden represent annual deposits, De Geer realized that if individual varves could be identified in two or more localities they might be correlated throughout a broad region and used as a measure of time.[26] On this basis he began studies in 1904 which led, 36 years later, to the publication of a vast definitive report on his results.[27]

De Geer's correlations are based fundamentally on the thicknesses of individual varves relative to varves lying above and below them. It is the relative thickness that is important, because the absolute thickness of a varve varies from place to place according to its position with respect to the source or sources of the sediment of which it is built. Vertical exposures of varved sediments in clay pits, river bluffs, and other excavations are cleaned to smoothness with spade and trowel, and sections disturbed or contorted by sliding or other agencies are avoided if possible. The top and base of each varve, and the position of the gradational contact between the coarse member and the fine member, are marked off on a narrow strip of paper held vertically (Fig. 72).

These field measurements are later plotted as curves on graph paper, the number of varves in the measured section being plotted against the

[26] De Geer 1912.
[27] De Geer 1940.

thickness of each varve (Fig. 72). Two curves constructed in this way from exposures in different localities are then placed side by side and one is moved up or down with respect to the other until, by this process of trial and error, a good fit or match is obtained (Fig. 73). Then, if

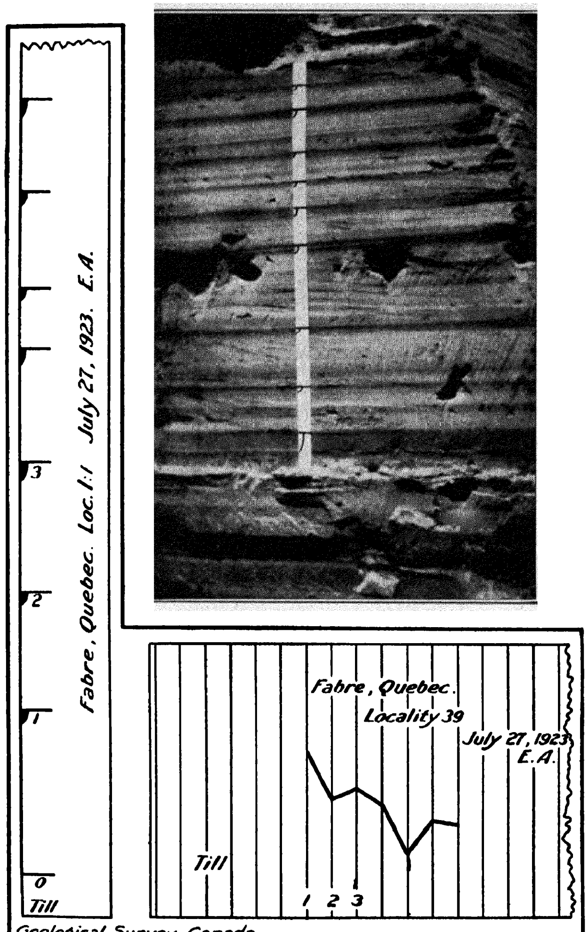


FIG. 72. The photograph shows the method of measuring varves on a strip of paper. The curve represents the thicknesses of seven consecutive varves, constructed from a paper strip (left) similar to that in the photograph. (Antevs.)

Varve 23 at Locality *A* is considered to match Varve 56 at Locality *B*, these are taken to represent the same year, and a composite chronology applying to these two localities and to the area between them is considered to be established. Varve 71 at Locality *B* may be found by this method to match Varve 3 at Locality *C*, and the composite correlation,

This correlation based on varve thickness rests upon the assumption that during the episode of deposition the weather varied from year to year and that the weather characteristics of any year (especially the summer season) were reflected, through the amount of glacier ice melted, in the amount of sediment deposited in the glacial lake or lakes throughout a region whose width equals the distance between the two mutually most distant localities in which the same single varve is recognized.

De Geer early found empirically that between curves obtained from localities only 0.7 mile apart he could usually get good fits. Hence he felt

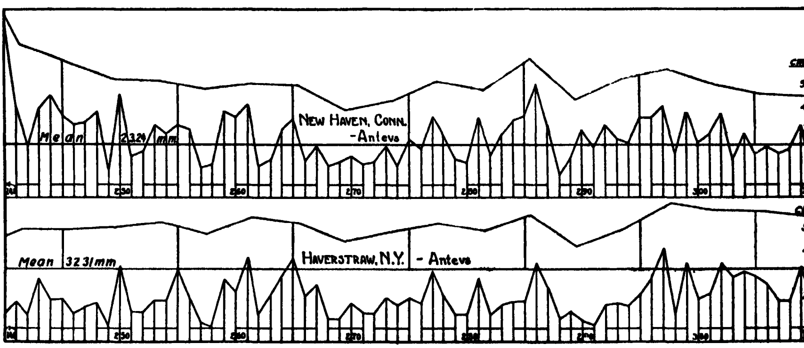


FIG. 73. Curves constructed from seventy consecutive varves measured at New Haven, Connecticut (above), and from seventy varves measured at Haverstraw, New York (below). The two curves have been placed in the position in which they are believed to correspond. (Reeds.)

confidence in extending his measurements through a distance of about 600 miles, from Scania, in southeastern Sweden, to Ragunda, in central Sweden, the measurements being, as nearly as practicable, 0.7 mile apart. He found that individual varves could be recognized in the curves through distances up to 35 miles, and that the varves lapped off upon each other like shingles on a roof, extending progressively farther north with increasing height in the section. When this work was completed and extended still farther southward, De Geer and his students had compiled a chronology, which, based on De Geer's correlations of the curves obtained, embraced about 8000 years as the time involved in the retreat of the margin of the Scandinavian Ice Sheet from the southern tip of Sweden to Ragunda and the deposition of lacustrine silt and clay.[28]

To this Lidén added a later, so-called postglacial period of about 8700 years, but even so he did not succeed in connecting this with historic time.

[28] De Geer 1929, p. 689.

The sequence of varves as worked out by De Geer and his students is not continuous; there are gaps in it across which the chronology was interpolated on a basis of inferences drawn from related eskers.[29]

De Geer also attempted to work out the *rate* of deglaciation by determining the headward limit of each overlapping varve. In order to do this it is necessary that the bottom varve in each measured section be exposed. According to De Geer most of the 700 closely spaced exposures in the Stockholm district reached the bottom varve, and from them a rate of retreat of several hundred feet per year was inferred. Farther north the rate is inferred to have increased, but throughout the Swedish record it is believed to have been of this general order of magnitude.

Later Antevs applied the method of varve measurement to the glacial-lake sediments of eastern North America.[30] Owing to the presence of serious gaps in the record, his first conclusion was that "the time elapsed since the ice sheet disappeared from New England can not even be estimated with any claim to accuracy."[31] Later he estimated the time involved in the retreat of the Wisconsin ice-sheet margin from Long Island to Cochrane, Ontario (about midway between the Great Lakes and Hudson Bay), at about 28,000 years, and the time from the Long Island position to the present at 40,000 (± 10,000) years.[32] Of this total number of years about 19,000 years are based on varve measurements; the remainder are interpolated, principally across three gaps. These are: Long Island to Hartford, Connecticut, about 50 miles; St. Johnsbury, Vermont, to Stony Lake, Ontario, about 120 miles; Mattawa, Ontario, to Lake Timiskaming, Ontario, about 170 miles. In addition, at Claremont, New Hampshire, there is another gap which Antevs assumed amounted to 200 or 300 varves.[33]

During the progress of Antevs' work in North America, De Geer extended his varve analysis to include the correlation of varve curves from localities far from each other. He correlated curves, for example, between Sweden and North America and between Sweden and South America, inferring that the same groups of years are identifiable on these pairs of curves, and that thus an absolute intercontinental correlation is created.

A varve is a stratigraphic unit, and like other stratigraphic units it can be followed throughout its extent only if its exposure is continuous. The normal procedure of tracing a layer of bedrock between exposures

[29] De Geer 1912, p. 248.

[30] Antevs 1922; 1925*b;* 1928*a;* 1931.

[31] Antevs 1922, p. 48.

[32] Antevs 1928*a*, pp. 155, 168; data elaborately recalculated by Bryan and Ray (1940, pp. 58–67).

[33] Antevs 1922, p. 83.

involves the use of some distinctive recognizable character in either the lithology or the fossil content of the stratum. But a given clay-silt varve is likely to differ from those immediately above and below it, if it differs at all, only in its thickness. Furthermore, the absolute thickness of a varve decreases away from the source of sedimentation; hence the distinctive character, if any, of a varve is not absolute thickness but relative thickness. In order to identify a given varve in two different exposures, one near and the other far from the source of supply, it must be assumed that the relative thickness of all the varves in the section remains the same, in other words, that each varve thickens headward at a proportional rate. Further, in districts where the surface beneath the varved sediments is sharply irregular, the thickness of a single varve may be greatly variable, thinning on steep slopes and thickening on gentle ones. This is true in the Stockholm district, but because the exposures there are very close together De Geer felt great confidence in his correlation despite the variations he had to deal with.[34] His method in such situations was to use the thickness variations of distinctive groups of varves rather than of individual varves, although, as he admitted, the groups as such are not everywhere normally developed.[35]

Few would doubt their own ability to correlate two varve curves from exposures, say, 200 yards apart in a single body of sediment, allowing their results to be checked by artificial completion of the exposure across the intervening distance to prove the correctness of the correlation by continuous tracing. On the other hand many have doubted the validity of De Geer's intercontinental correlations involving pairs of localities many thousands of miles apart.[36] The question of the reliability of varves as correlation units is therefore largely the question: through how great a distance can two varve curves be correlated with confidence? Different persons experienced in varve study have given different answers to this question. De Geer went farthest in connecting curves through great distances. An excellent critical appraisal of his method of intercontinental correlation was published by Antevs, who concluded that it is unreliable.[37] An interesting reply was made by De Geer.[38]

Antevs' point of view was more conservative than that of De Geer. He said:

> Agreement between curves is determined by the trend of lines connecting points denoting, in this case, the thicknesses of the annual

[34] De Geer 1934, p. 2.

[35] De Geer 1934, p. 13.

[36] Cf. Brückner 1921; Antevs 1922, p. 48; 1931, p. 35; Reeds 1929, p. 624; Ellsworth and Wilgus 1930; Coleman 1941, pp. 86, 153.

[37] Antevs 1931, pp. 34–41.

[38] De Geer 1934.

> deposit. The transportation and sedimentation of mud in a glacial lake being influenced by a number of conditions, such as shape of the basin, islands, currents, and the chemical and physical properties of the glacial mud and of the lake water, a perfect agreement cannot be expected between varve graphs. The degree of correspondence that shall be deemed necessary for correlation is, therefore, left to the correlator's discretion. Because there is no simple way to determine or objectively to estimate the degree of actual correspondence between two graphs, it follows that the appreciation of the similarity is also subject to personal judgment. Varve graphs are individually only moderately distinctive, and, in addition, being records of the summer heat, they present periodic, more or less similar fluctuations.
>
> These conditions make it necessary to exercise considerable conservatism in assuming varve correlations. Correlations cannot be made on agreement of curves alone. It is necessary to know from striae, moraines, eskers, drumlins, and other features, that the compared varve series are of approximately the same age.[39]

Sauramo was even more conservative than Antevs. He stated[40] that correlation based exclusively on varve thickness may lead to unreliable results, for three reasons:

1. In many sections the variation in thickness of adjacent varves is so small that the curve constructed from them approaches a straight line and precludes correlation.

2. Thicknesses commonly vary in a horizontal direction so that the same sequence of varves appears different in curves from two or more localities; hence correlation can not be made. This occurs very commonly and may be true for two localities very close together, as Sauramo showed. On the other hand, he agreed that under favorable conditions individual varves maintain their thicknesses relative to each other and hence can be correlated from place to place.[41]

3. Correlations based on the matching of two curves by this method have been shown, on the basis of the stratigraphic relations of the sedimentary bodies concerned, to be wholly erroneous.[42]

Sauramo might also properly have indicated that there is disagreement or uncertainty in some localities as to whether the measured units are of annual or lesser value.[43]

In consequence, Sauramo studied varves as the stratigrapher studies layers of rock devoid of fossils: he made a sedimentologic and strati-

[39] Antevs 1931, p. 36.
[40] Sauramo 1923, p. 11.
[41] Sauramo 1929, pp. 41, 52.
[42] Sauramo 1918, p. 34.
[43] Cf. Antevs 1931, p. 31; Fraser 1929; Speight 1926, p. 71.

graphic study, grouped the individual varves into varve series, recognized key strata identified by color, grain size, and other physical properties, and noted facies changes. He made no attempt to correlate individual varves until the varve series had been identified. He said:

> By thus making *first* the correlation of *larger*, and then of *successively smaller* units of sedimentation, the probability of error is reduced as far as possible. Each connection is checked from several sides, being based upon not one but several characters which are in part independent of each other. It is then always possible to tell whether the connection has reached the accuracy of single varves or whether the possible error remains larger, and how large it is. Connection based upon variation in thickness only affords no check of this kind, and the amount of possible error can not be estimated at all.[44]

Even the varve correlation made by De Geer and Antevs through the very short distance between Denmark and southern Sweden was severely criticized on the ground that the implied relative dates of the several Danish deposits concerned are in complete conflict with the stratigraphic evidence.[45]

The whole matter of the reliability and usefulness of varve correlation is at present in an unsatisfactory state. Largely because it has been subjected to an inadequate amount of criticism and discussion, most geologists have no definite opinions on it. Nor are they likely to have opinions of value until varve correlation has been objectively tested by a number of investigators working independently on the same exposures. As Antevs pointed out, correlation is to some extent a matter of judgment. The extent to which judgment plays a part is capable of objective testing, and, until these tests are applied, varve study is not likely to progress far beyond its present state. Meanwhile no one familiar with varve literature can fail to be impressed by the half century of tedious groundwork accomplished by De Geer and his helpers, or by the enthusiasm with which varve studies have been carried on.

GLACIAL AND INTERGLACIAL CHRONOLOGY

A few attempts have been made to estimate the lengths of various glacial and interglacial subdivisions of the Pleistocene epoch. All but one of them are based on comparison of the progress reached by some process, such as weathering or general erosion, during some stratigraphically known time, with the progress reached by the same process in postglacial time within the same region. Assuming an absolute value

[44] Sauramo 1923, p. 13.
[45] Milthers 1927.

for postglacial time, they are thus able to convert the estimate into absolute terms. We have seen, however, that the absolute duration of postglacial time has not been determined for any region with even an approach to accuracy; in consequence the estimates of earlier Pleistocene time units include this inaccuracy as well as others inherent in the bases used for the estimates. If the values estimated for postglacial time in various districts are at least of the right order of magnitude, then perhaps the same can be said of the more successful estimates for the earlier times, but, as all of them are largely matters of judgment, the reader must form his own opinion of the merits of each from the brief summaries that can be given here and from the original studies on which the summaries are based.

Inference from Degree of Weathering

Pedestals Beneath Erratic Boulders

Matthes estimated the time since the El Portal glaciation in Yosemite National Park, California (tentatively correlated with the Illinoian glaciation), in the following manner:[46] The summit of Moraine Dome was glaciated during the El Portal Glacial age but not during the Wisconsin Glacial age. On this summit a glacial boulder is perched on a pedestal 7 feet high and 4 feet thick. The pedestal consists of the remains of an aplite dike which, being resistant to weathering, has developed the pedestal while the surrounding quartz monzonite, a weaker rock type, has disintegrated. As the dike pedestal is too thin and delicate to have withstood glaciation, it must entirely postdate the El Portal age. In contrast, the same kinds of rocks in this vicinity that were glacially smoothed during the Wisconsin age are so fresh that they show hardly any weathering at all. Hence the El Portal glaciation is estimated to be ten to twenty times as remote in time as the Wisconsin glaciation. On the assumption that the freshly glaciated rock surfaces in this vicinity were abandoned by the Wisconsin ice 10,000 years ago, the end of the El Portal (Illinoian?) Glacial age here would then be placed at 100,000 to 200,000 years ago.

This estimate makes no pretense to accuracy, but on the facts given it probably does represent the order of magnitude of the elapsed time since the earlier glaciation.

Decomposition of Pleistocene Sediments

Depth of leaching in Iowa. A body of drift begins to be decomposed by atmospheric agencies, aided by subsurface water, as soon as it is

[46] Matthes 1930, pp. 70–72.

exposed. After decomposition has progressed for a long time, it produces a zonation of the drift (whether till, gravel, or other material) as follows:

Surficial soil.
Chemically decomposed till, consisting chiefly of alteration products and of the most resistant constituents of the original drift.
Leached and oxidized drift.
Oxidized drift.
Unaltered drift.

The arrangement of these zones shows that, as might be expected, oxidation progresses more rapidly than leaching, and that leaching takes place more rapidly than thoroughgoing decomposition of the drift.

By examining a very large number of drift sections in Iowa, Kay found that the "depth of leaching," that is, the distance from the surface down to the base of the leached and oxidized zone, is very nearly uniform on any drift sheet regardless of whether the material consists of till, or gravel, or loess, or any combination of them, provided only that the sections measured are situated on broad interfluves with very little slope.[47] He reasoned that depth of leaching is a fairly good indicator of the time during which decomposition was in progress. Where a drift sheet is overlain by another drift sheet, this time is equal only to the length of the intervening interglacial interval; where no such cover exists, the time of decomposition runs right up to the present.

Kay found that in the Mankato drift the average depth of leaching is 2.5 feet, whereas in the Iowan drift, where not covered by later deposits, it is 5.5 feet or 2.2 times the depth in the Mankato. Assuming post-Mankato time in Iowa to equal 25,000 years, he multiplied this figure by 2.2 in order to obtain an estimate for post-Iowan time. Applying this method to the earlier drift sheets, he derived figures that may be summarized as in the tabulation on page 400.

The absolute value of these figures depends on the initial assumption as to the length of post-Mankato time. Their relative value depends to some extent on whether or not the rate of leaching is a straight-line function of depth, at least down to a depth of 30 feet — a factor carefully noted by Kay. If it is not, then the absolute value of each result must be increased, perhaps greatly, and the relative values of the group will differ from those given.

The combined lengths of the three recognized interglacial ages, plus all post-Iowan time, according to these estimates total 675,000 years. But this figure does not include the Nebraskan, Kansan, and Illinoian glacial ages, or the Iowan sub-age of the Wisconsin age, there being no comparable basis for estimating the lengths of these units.

[47] Kay 1931.

Type of Material	Stratigraphic Position	Depth of Leaching (feet)	Factor	Result (years)
Gravel	Wisconsin (Mankato)	2.5		25,000 (assumed)
Gravel	Iowan (never covered by younger deposits)	5.5	2 2	55,000 (measures all post-Iowan time)
Till	Illinoian (where covered by Iowan drift)	12 0	4 8	120,000 (measures Sangamon Interglacial age)
Till	Kansan (where covered by Illinoian drift)	30 0	12 0	300,000 (measures Yarmouth Interglacial age)
Till	Nebraskan (where covered by Kansan drift)	20 0	8 0	200,000 (measures Aftonian Interglacial age)

Depth of leaching in Indiana. Thornbury made similar estimates for drifts in Indiana, following in general the method of Kay.[48] Assuming a value of 25,000 years for post-Mankato time, he estimated the Cary-Mankato interval at 10,000 years, and the Tazewell-Cary interval at 10,000 years, making the values of post-Cary and post-Tazewell time 35,000 years and 45,000 years respectively. He found the depth of leaching of Illinoian drift in Indiana, where covered by Wisconsin drift, to be only 6 feet, whereas Kay had found this depth to be 12 feet in Iowa. Thornbury attributed the difference to lower relief and poorer drainage in Indiana than in Iowa, during Sangamon time. However, by using the rate of formation of gumbotil (estimated at 1 foot in 19,000 years following an initial period of 50,000 years during which no gumbotil forms) instead of depth of leaching, Thornbury estimated that Sangamon time endured 135,000 years, as against Kay's estimate of 120,000 years based on depth of leaching.[49] Both methods, of course, are based ultimately on an assumed value of 25,000 years for post-Mankato time.

Thicknesses of residual soils in Bermuda. The stratigraphy of the island of Bermuda consists chiefly of calcareous eolianites (limestones originally deposited by the wind), alternating with red and brown clay soils. Each of the soil layers is the product of protracted chemical decay of the eolianite that underlies it, and the decay of each took place before the next overlying eolianite was deposited. The eolianites are believed

[48] Thornbury 1940.
[49] Thornbury 1940, pp. 473–474.

to have been accumulated during the glacial ages, when sealevel was low, and when winds were strong owing to displacement of the polar-front zone toward the south. The soil-making times are thought to have coincided with the interglacials, when sealevel was high and when winds were gentler, like those of today. The use of these soils in an attempt to calculate the values of Pleistocene time units we owe to Sayles.[50] Sayles started with the assumption of Kay, already referred to, that "postglacial" (actually post-Mankato) time has endured 25,000 years. As the soil now forming on flat surfaces on the stratigraphically highest eolianite in Bermuda is approximately 6 inches thick, and as this soil is believed to postdate the last glaciation (though no attempt was made to define its relation to the Wisconsin sub-ages recognized in North America), it was inferred that the time required to create 1 foot of soil on a flat surface is 50,000 years. (Incidentally 1 cubic foot of soil is calcu-

Soil	Thickness (feet)	Estimated Period of Development (years)	Suggested Correlation with North America
Soil now forming	0.5	25,000 (assumed)	
Eolianite			
McGall's soil*		10,000 ?	
Eolianite			Wisconsin
Signal Hill soil*		10,000 ?	
Eolianite			
St. George's soil	2.0	120,000	Sangamon
Eolianite			Illinoian
Harrington soil	2.0		
Marine limestone†		250,000	
Shore hills soil	2.0		Yarmouth
Marine limestone†			
Eolianite			

* Found exposed on slopes only; probably developed rapidly, in part by slopewash.

† These layers necessarily represent high sealevels and hence are referable to interglacial rather than glacial ages.

[50] Sayles 1931, pp. 452–456.

lated to require the solution of 100 cubic feet of eolianite.) Applying this figure to the measured thicknesses of the various soil layers, Sayles arrived at the figures shown in the tabulation on page 401.

This calculation encounters the same fundamental difficulty that Kay encountered: the basic figure, 25,000 years, has to be assumed. However, the relative magnitudes of the figures are significant, especially when compared with the figures derived by Kay. The importance of determining with accuracy the length of postglacial time for each of several regions becomes more and more obvious, although for Bermuda this would be difficult because, even on the assumption that the eolianites record low sealevels and the soils high ones, there is as yet no means of correlating the beginning of formation of the modern soil in Bermuda with any particular event recorded by the stratigraphy of North America.

Inference from Degree of Erosional Loss

An estimate based on the degree of erosional loss by the several drift sheets in the Mississippi basin was attempted by Leverett.[51] He too began with the assumption that post-Mankato time has lasted 25,000 to 30,000 years. On this basis he guessed that all post-Tazewell time must be about 70,000 years long.

Leverett then estimated the average thickness of the prism of drift that has been removed from each of four drift sheets in districts where they have not been covered by younger deposits. Taking as a basis his guess at the length of post-Tazewell time, he then calculated the lengths of post-Illinoian and post-Kansan time, as follows:

Stratigraphic Unit	Erosional Loss (feet)	Time since Uncovered by the Ice
Tazewell	5	70,000 (assumed)
Illinoian	15	210,000
Kansan	50	700,000

Comparison of the figures with those arrived at by Kay and by Thornbury shows that Leverett's post-Tazewell figure is 25,000 years longer than Thornbury's, and 15,000 years longer than Kay's estimate for the whole of post-Iowan time. If we use Thornbury's estimate of 45,000 years for post-Tazewell time, and follow Leverett's method, we get 135,000 years for all post-Illinoian time, whereas Thornbury gets 135,000 years for the duration of the Sangamon age alone. Both methods apply the same (assumed) yardstick, namely, 25,000 years for post-Mankato time; hence the discrepancy lies in the difference between comparative depth of leaching and comparative degree of erosional loss. The former should be

[51] Leverett 1930*b*.

the more reliable because it is subject to fewer variable influences and also because it is less difficult to measure accurately. Leverett's unparalleled experience in the observation of drifts undoubtedly enabled him to estimate their erosional losses more closely than anyone else, but the fact remains that the losses had to be estimated because they could not be measured.

Holmes described an interglacial valley at Chittenango Falls State Park in central New York.[52] The valley was filled with drift during the latest glaciation of the region and has since been partly re-excavated. The interglacial valley itself is 8.5 times as long as the postglacial excavation. Holmes assumed the duration of postglacial time here to have been 20,000 years, and he assumed further that the rates of lengthening of the two cuts were comparable. Thus he arrived at 170,000 years as the time required for the cutting of the interglacial valley. The method is necessarily crude but gives results of an order of magnitude comparable with those inferred from the gumbotils.

Deduction from Astronomic Relations of Earth to Sun

If the climatic changes responsible for the glacial and interglacial ages were proved to be controlled by astronomic changes that are periodic, the periods of the changes being known, then it would be possible to date accurately the principal glacial and interglacial events. Three periodic or near-periodic changes in the elements of the Earth's orbit are known. These affect (1) the eccentricity of the orbit, with a period of about 91,800 years; (2) the angle between the Earth's axis and the plane of the ecliptic, with a period of a little more than 40,000 years; (3) the shifting of the perihelion (causing precession of the equinoxes), with a period of slightly more than 21,000 years. Each of these changes affects the amount of solar radiation received by any given part of the Earth throughout its period. When the three kinds of changes are combined, instead of being considered separately, it becomes possible to construct for any latitude a curve showing (subject to errors introduced by variable factors) the amount of solar heat received throughout a long period of time.[53]

Milutin Milankovitch calculated such a curve for latitude 65° N, using 176 dates as control points and considering only the summer months, because summer temperatures are more critical for glacial fluctuations than winter temperatures. The curve (Fig. 74) is of course

[52] Holmes 1935.
[53] See the discussion of this matter in Chapter 22 and the references there cited.

irregular, because it combines three different influences each having a different period.[54]

The curve shows nine conspicuous radiation minima arranged in four groups of two or three minima each. These four groups are regarded by some authorities as representing the four glacial ages recognized in the region of the Alps by Penck and Brückner, and the two or three minima within each group are compared even more closely with elements in the geologic record, by some who profess to see in the curve a fairly complete and accurate picture of the climatic changes of the Pleistocene epoch.[55]

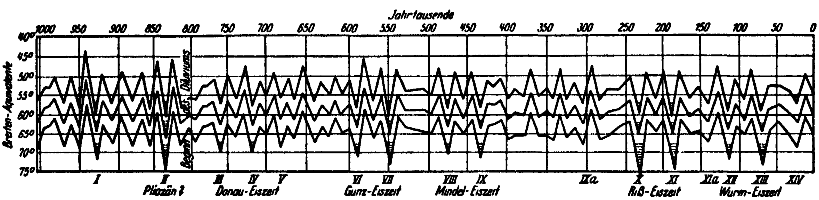


FIG. 74. Curves showing deduced variations in solar radiation received at 55°, 60°, and 65° north latitude during the summers of the one million years preceding A.D. 1800. Horizontal scale = units of 50,000 years. Vertical scale = latitude in 5-degree units. Thus "latitude-equivalents" — imaginary changes of latitude — are shown instead of changes in the actual amount of radiation received at one latitude. (Milankovitch.)

On the assumption that the Milankovitch curve actually represents the geologic record, the absolute date of any major glacial or interglacial event can be fixed by deduction from the curve. The following table, greatly modified from Zeuner,[56] shows the dates of major events as thus interpreted:

STRATIGRAPHIC SUBDIVISION	DATE OF MAXIMUM (years before 1800 A.D.)
Wurm 3 (= Pomeranian)	18,000
Wurm 2 (= Brandenburg)	67,000
Würm 1 (= Warthe)	112,000
Riss/Wurm Interglacial	143,000
Riss 2 (= Saale)	183,000
Riss 1	226,000
Mindel/Riss Interglacial	
Mindel 2 (= Elster)	430,000
Mindel 1	472,000
Günz/Mindel Interglacial	
Günz 2	545,000
Günz 1	586,000

[54] Milankovitch 1920; 1930; 1938.
[55] See especially Soergel 1937; Zeuner 1935.
[56] Zeuner 1935, p. 357.

This plausible scheme incurs several serious objections, which are summarized in Chapter 22, as are the similar schemes put forward by Spitaler and by Köppen and Wegener. These objections are so serious that this method of establishing a chronology is of very doubtful value. The method can hardly be applied with confidence until its proponents have disposed of the evidence against it.

Inference from Concentration of Radium in Sea-Floor Sediments

A different contribution to the determination of an absolute chronology of Pleistocene events was embodied in the work of Piggot and Urry on the concentrations of radium in sea-floor sediments.[57] The helium and lead methods applicable to rocks of greater age are not suited to the determination of dates as recent as the Pleistocene. However, these investigators used radioelements in an entirely different manner. They found that the content of radium and other radioactive elements in marine sediments varies with time in a very complicated way. On this basis they established a standard curve of radium concentration plotted against time. By comparison with this curve the age of a layer of sediment (that is, the elapsed time since its deposition) can be determined. Unfortunately the radium content becomes essentially constant after 300,000 to 400,000 years. Hence the method is suitable only for times within this limit and is not applicable to the whole of the Pleistocene epoch.

Thus far the radium-concentration method has been applied only to cores taken from the North Atlantic sea floor, where, however, its accuracy has been checked and found satisfactory by reference to a "key horizon," a distinctive layer of volcanic ash recognizable in the cores. The potential value of this method for dating events during at least a part of the Pleistocene is great.

SUMMARY

Examination of the data on glacial and interglacial chronology reveals the fact that absolute dating of Pleistocene events has not yet been accomplished. Most of the methods of dating glacial and interglacial events take as a baseline a postglacial span of time that has to be assumed. Much ingenuity has been expended on the problem, but the problem still remains unsolved. It is possible that the best estimates of postglacial time for any region are of the right order of magnitude, but, as we go farther back into the Pleistocene, the lengths of time involved

[57] Piggot and Urry 1942.

become rapidly more uncertain. The length of the whole Pleistocene epoch appears to be not less than a million years; certainly not more than five million and probably not more than two million years. We may well doubt that data we have been able to gather thus far justify any closer approximation. On the other hand, studies of radium concentration in sea-floor sediments may lead to the establishment of a satisfactory chronology of the last 300,000 years.

The dates shown in Table 30 are no more than estimates.

Chapter 19

CHANGES IN LEVEL OF LAND AND SEA

INTRODUCTION[1]

The building and wasting away of vast ice sheets have the direct effect of causing changes of level of two distinct kinds: warping of the Earth's crust, and fluctuation of the level of the sea. The first of these changes, crustal warping, affects only the regions actually covered by and peripheral to the ice sheets and is caused by the weight of the ice, which constitutes an extra load on the crust. If a glacier is very large, its weight is great enough to overcome the strength of the weak, easily deformed[2] substratum immediately below the stronger outer crust. The crust subsides in basinlike fashion beneath the ice, and in the substratum plastic flow transfers rock material outward away from the basined area, in compensation for the extra load caused by the glacier. When the glacier begins to shrink, the reduction in weight causes a deficiency of load; the direction of plastic flow in the substratum is reversed, the crust bulges up in domelike fashion,[3] and after the glacier has disappeared entirely — though it is some time afterward, owing to a marked time lag between removal of load and completion of subcrustal flow — conditions are restored to normal.

Indirectly because of this time lag it becomes possible to measure the amounts and rates of uplift at various places. Along the seacoast, and in regions where the shrinking glaciers were fringed by lakes of glacial meltwater, extensive areas were submerged as soon as they were uncovered by the ice. The waves and currents of the sea and of these lakes fashioned cliffs, beaches, and the like along the shores, thus creating horizontal strandlines. The lagging upwarping of the crust throughout the regions vacated by the ice, however, caused wide areas to emerge from beneath the water, and with this emergence the strandlines sculptured in the temporarily submerged areas were lifted up and warped

[1] A. C. Lawson 1940 contains a very clear general statement of the relation of ice sheets to crustal warping.

[2] The deformation is chiefly plastic, but elastic distortion takes place also.

[3] The bulge is domelike structurally but it does not of course necessarily form a topographic dome. An area essentially horizontal to begin with would, after glacial subsidence and subsequent upwarping, be essentially horizontal again.

out of their original horizontal positions. Those which are still preserved are therefore inclined, in general standing highest where the former glacier ice was thickest and where the postglacial updoming was greatest. The inclination of a strandline thus warped out of its original horizontality measures the differential uplift, in the direction of trend of the strandline, that took place between the time when the strandline began to rise above the water level and the present time. Because of the time lag between deglaciation and crustal recovery, and because in some regions several successive strandlines are warped in this way, it has been possible to form a good though incomplete picture of the rate and progress of postglacial doming both in northwestern Europe and in eastern North America. The doming movement began so recently that it is still in progress.

The second change of level that results directly from glaciation is not confined to the glaciated regions but is worldwide. It consists of the fall and rise of sealevel throughout the world. The building of large glaciers when none existed before requires that vast quantities of atmospheric moisture be precipitated in the form of snow. This moisture must come ultimately from the sea. The evaporation of sea water is of course constantly taking place, but normally the water thus evaporated is precipitated chiefly as rain and is returned to the sea through rivers and other streams as rapidly as it is evaporated. Under glacial climates, on the other hand, the proportion of snowfall to rainfall increases, and the precipitated snow, converted into glacier ice, is returned to the sea so slowly and sluggishly that at any one time large amounts of it are locked up, as it were, on the lands. The result is lowering of the sealevel.

Under nonglacial climates the existing glaciers melt and large volumes of water are returned to the sea, thereby raising its level. The alternation of glacial and interglacial ages has resulted in fluctuation of the sealevel through a range of some hundreds of feet, including levels both higher and much lower than the level with which we are familiar today.

There is evidence, much of it fragmentary and largely in the form of marine benches, strandlines, and fossil-bearing deposits, both of the low sealevels of the glacial ages and of the high sealevels of interglacial times. Some glaciated coastal regions carry evidence of both postglacial upwarping of the crust and postglacial rise of sealevel. Usually such evidence is difficult to sort into its two component parts, but it is important at least to be able to recognize that in a given region both kinds of change have taken place.

CRUSTAL WARPING

Direct Relation to Glacial Loading and Unloading

Formerly there was some question whether the upwarping of abandoned shorelines and the continuing present uplift in northern Europe and North America[4] were caused by deglaciation. It was held by some that this warping was unconnected with glaciation and that it was merely a subordinate continuation of the many gentle warping movements known to have affected wide areas of these continents repeatedly throughout the Cenozoic Era. There is no longer any doubt, however, that the updoming in question is the direct result of the removal from the crust of the excess loads consisting of two vast ice sheets. The supporting evidence, essentially as summarized by Gutenberg,[5] is as follows:

1. In both Fennoscandia and North America the outer limit of the upwarped region parallels the limit of the latest glaciation, and in both regions it has remained in the same location through the last several thousand years.
2. In both regions the isobases (lines connecting points on any strandline that have been equally uplifted) are concentric to the area in which, according to independent evidence, the former ice was thickest and therefore heaviest.
3. In both regions the rate of uplift is of the same order of magnitude. During the last few thousand years the rate of uplift of Fennoscandia has slowed down from a rate about twice as great, but the available evidence does not yet justify a similar statement about North America.
4. In both regions even the incomplete data obtained thus far show that gravity anomalies are negative,[6] and that they increase toward the central parts of the glaciated areas, thus indicating that crustal equilibrium in these regions has not yet been reached.
5. In other regions of former glaciation, or former greater glaciation, such as Great Britain, Greenland, Spitsbergen, Novaya Zemlya, Siberia, Patagonia, and Antarctica, where postglacial upwarping is expectable, evidence of it has been observed.

So clearly is the direct relation between large glaciers and crustal warping now recognized that the presence of late-Pleistocene emerged marine deposits occurring widely throughout a high-latitude region

[4] The effect of the Siberian Ice Sheet on crustal warping is as yet unknown.

[5] Gutenberg 1941, p. 750.

[6] That is, there are deficiencies of mass in these sectors of the globe. When in future the plastic return flow in the substratum and the accompanying upwarping of the crust have been completed, presumably these deficiencies will have been eliminated.

may be taken as indicating (although it does not prove) the probability of extensive glaciation. This line of reasoning has been applied by Washburn[7] to parts of Arctic Canada where direct evidence of glaciation is scanty or has not yet been found.

Before the principles of crustal equilibrium had become well established, and when the crust was still believed by some to be so rigid that it could sustain the extra loads created by the ice sheets without bending, it was suggested that the inclined postglacial shorelines in the glaciated regions were the result of the distortion of lake levels and sealevel through gravitative attraction by the great masses of the ice sheets. However, comparisons of detailed quantitative calculations such as those of Woodward[8] have shown that, on the most favorable assumptions as to the masses of the ice sheets, the resulting distortion of water levels would be only a fraction of the measured inclinations of strandlines. But not even this fraction is actually attributable to gravitative attraction of the glaciers, because subsidence of the crust beneath the weight of the ice reduces the mass effective for distorting water levels. In consequence this factor is minor if not negligible.

An important question concerns the thickness of ice required to push down the crust. The answer is not known exactly, but we can say in general that there is no evidence that small glaciers have thus affected the crust. The small former ice cap of The Faeroes and that of Kerguelen Island in the southern Indian Ocean did not apparently bend their floors, although the larger, composite British ice caps did so. According to Daly[9] those former ice caps which are shown to have caused crustal subsidence were more than 300 miles in diameter and more than 3300 feet thick.

The amplitude of the crustal warping produced by ice-sheet loading depends on at least four factors: density of the ice, thickness of the ice, density of the crustal material beneath it, and degree of crustal adjustment to the additional load. Of these factors only the first is known with any degree of accuracy; the density of glacier ice, although variable, is not far from 0.9. The other factors must be estimated. The density of the crustal material transferred by flow is believed to be in the neighborhood of 3.0 to 3.3. Adopting these values and assuming that adjustment is complete, we can say that the maximum subsidence beneath a large ice sheet should be equivalent to one-third the maximum thickness of the ice, or a little more than one-third, according to the value adopted for the density of the crustal material.

[7] Washburn 1942, p. 150.
[8] Woodward 1888, pp. 60–79.
[9] Daly 1938, p. 182.

A related question is: what effect does strong topographic relief have upon the form of the warped surface? Do highlands and deep depressions introduce irregularities into the warping? Apparently their effect is small at most, for the isobases constructed upon warped strandlines appear to follow broad regional trends regardless of local topography. This fact, and also the failure of small glaciers to bend the crust, indicate that isostatic adjustment for differences of load does not take place locally but is distributed beneath a wide area.

Updoming in Fennoscandia

In two parts of the world studies of postglacial warping of the crust have been correlated across regions wide enough to give a general picture of the form of the warping—Fennoscandia and the eastern half of North America. That postglacial warping occurred in a number of other glaciated regions has been established, but the data on them are not yet sufficient for broad syntheses.

Inference from Warped Strandlines

The postglacial upwarping of Fennoscandia is established on two lines of evidence: (1) warped strandlines, and (2) tide-gage records showing that the land is still rising relatively to the sea. The strandlines, varying considerably in distinctness from place to place, were fashioned by the water bodies that successively occupied Fennoscandia during the shrinkage of the ice sheet: Baltic Ice Lake, Yoldia Sea, Rhabdonema Sea, Ancylus Lake, and Littorina (Tapes) Sea, described in Chapter 15. In Finland about thirty such strandlines have been recognized.

Isobases, essentially contours, constructed on the surface represented by any one strandline show that the subsequent warping of the crust has a domelike form. The central part of the dome revealed by each set of isobases lies in the general region of the Gulf of Bothnia, where the ice sheet was thickest. Figures 75, 76, and 77 show the isobases constructed by Sauramo for three of the later strandlines.

Whereas the younger strandlines extend throughout the Baltic region, the older ones are present only in the southeastern part of the region, because, at the times when the older ones were made, the rest of the region was still occupied by the shrinking ice sheet. Furthermore, as soon as the ice sheet began to be thinned, the crust began to rise. But actual shrinkage of the glacier margin had to take place, permitting lake or sea water to occupy part of the region formerly covered with ice, before

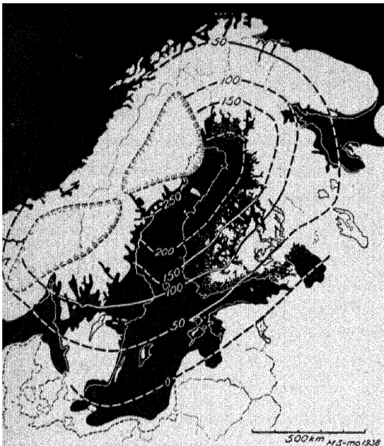


Fig. 75. Fennoscandia at the time of the first Rhabdonema Sea. (Sauramo.)

Marine areas are shown in black. The faint hatched lines indicate the two chief remnants to which the Scandinavian Ice Sheet had been reduced by this time. The curves are isobases on the strandline of the first Rhabdonema Sea. The interval between the curves is 50 meters. The curves show that this strandline has been domed up to positions as high as 250 meters above present sealevel.

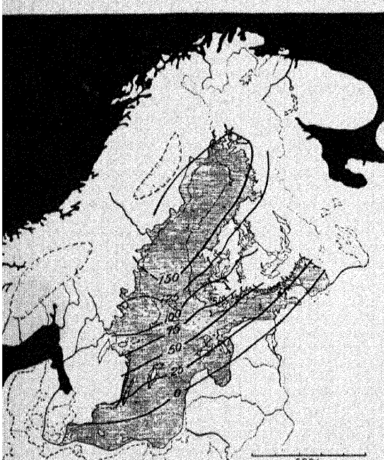


Fig. 76. Fennoscandia at the time of the first appearance of the Ancylus Lake. (Sauramo.)

All conventions are as in Fig. 75.

Fig. 77. Fennoscandia at the time of the beginning of the Littorina Sea. (Sauramo.)

All conventions are as in Fig. 75.

shoreline-making could start. Hence it is inevitable that some recovery of the crust by upwarping must have occurred before even the earliest strandlines were fashioned. As there are no means of measuring the amount of this early recovery, we are obliged to admit that the figure based on total measured recovery, a figure of the order of 275 meters, is a minimum, and that total actual recovery was greater than this by an unknown amount. For total actual recovery, estimates as high as 1500 meters have been made, but all are necessarily based on assumptions. Daly arrived at a figure of 600 meters.[10] Gutenberg calculated the figure at 500 meters.[11]

The maps (Figs. 75, 76, and 77) show that the isobases lie closer together near the center of the uplift than near its periphery. This spacing means that, translated from a map into a vertical profile, each restored water surface would appear concave up, steepening toward the center of uplift, and that therefore at all times the movement had the form of a differential, domelike bending.

That the updoming did not proceed at a uniform rate is shown by abrupt increases in the slope of a single strandline as viewed in vertical profile. Figure 78 shows in profile the more important strandlines of the Baltic region. Two sharp bends are shown. Each of these bends marks a discontinuity in the rate of uplift. It indicates that for some time after the shrinking ice had evacuated the district where the bend is now recognized the crust remained quiescent while the waves of lake or sea fashioned a single strandline. Later there set in a rapid uplift which affected only a part of the lake- or sea-covered region, and hence bent the strandline differentially. The episode of quiescence may have been caused by a climatic fluctuation that slowed down or even temporarily halted the shrinkage of the ice sheet.

The date of the increased bending can be readily fixed in terms of the strandline succession. Thus it is clear that the Finiglacial bend (*f*, Fig. 78) came into existence between the time when the shoreline of the Rhabdonema Sea (*Rha I*) was made and the time of the first Ancylus Lake (*A I*), because all the strandlines older than *Rha II* are bent whereas the younger ones are not. Further, regardless of localized bends, the amount of crustal warping that intervened between the times of any two water levels can be inferred directly by comparing the rate of slope of the younger strandline with the rate of slope of the older one. If this principle is applied to the profiles shown in Fig. 78, it is readily seen that the rate of uplift in Fennoscandia was not uniform.

[10] Daly 1934, p. 138.
[11] Gutenberg 1941, p. 751.

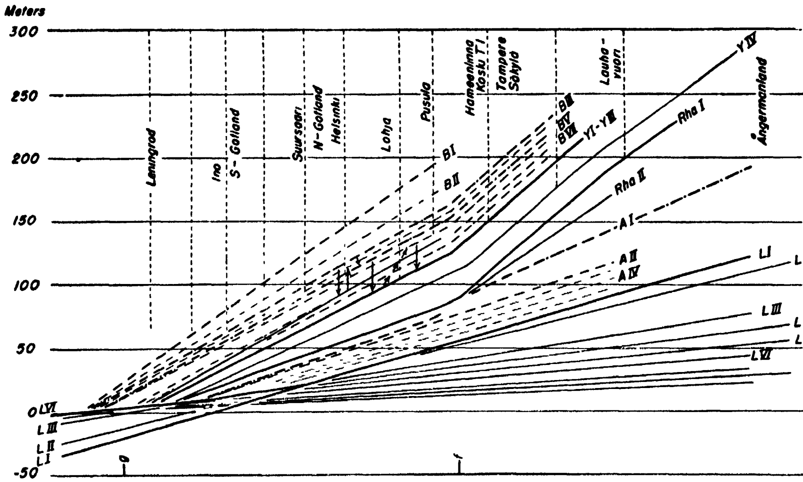


FIG. 78. Strandlines of the Baltic region seen diagrammatically in vertical profile from near Leningrad on the southeast (left), northwest to Ångermanland on the Swedish coast of the Gulf of Bothnia (right). (Sauramo.) Horizontal distance represented is about 500 miles.

Horizontal lines show altitudes in meters. Continuous oblique lines = sealevels. Dash-dot lines = lake levels. *B* = Baltic Ice Lake strandlines; *Y* = Yoldia Sea strandlines; *Rha* = Rhabdonema Sea strandlines; *A* = Ancylus Lake strandlines; *L* = Littorina Sea strandlines. The arrows indicate the chronologic sequence of strandlines between *B III* and *Y I*. *g, f* = bends in strandlines explained in the text.

Inference from Tide-Gage Records

Not only did the uplift of the crust take place during and immediately after the deglaciation of Fennoscandia; it is continuing at present. The form and rate of this continuing uplift are learned from the records of tide gages governmentally placed at many points along the shores of the sea and of large lakes as well. The gages are fixed firmly to the rocky shore and are provided with floats which show, and in some gages automatically record, changes in the position of the water surface with respect to the land. When the record of a large number of gages over a period of many decades is examined, and meteorologic and other extraneous effects are allowed for, it appears clearly that the domelike upwarping in Fennoscandia is still in progress. Although century-old casual observations of emerging rocks and shoals had made it certain that uplift was currently occurring, it was in 1918 that the inquiry was put on a sound and extensive quantitative basis by Witting.[12] His studies have

[12] Witting 1918.

been amplified by himself and by others since then, and the essential results have been summarized and critically discussed by Gutenberg,[13] who compiled a convenient chart (Fig. 79) in which the present rate of uplift is shown by curves drawn at intervals of 20 centimeters per century. From this chart it appears that the southern shores of the Baltic and the North Sea and the coast of Norway lie along the approximate outer limit of present uplift, and that the rate of uplift increases to more than 1 meter per century within a small central area at the head of the Gulf of Bothnia. The isobases are close together in Norway, farther apart in the Baltic region, and still farther apart in northern Sweden and the Gulf of Bothnia. This arrangement implies a broad, rather flat-topped dome with a steeply sloping margin,[14] a form slightly different from that implied by the isobases on the former strandlines already discussed.

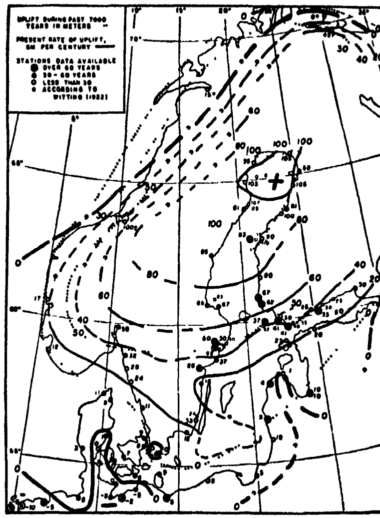


FIG. 79. Present rates of upwarping of the crust in Fennoscandia. (Gutenberg.)

Curves (broken lines) show present rate of uplift in centimeters per century and are plotted from tide-gage records. Compare the dotted-line curves, which show total uplift during the past 7000 years, as inferred from warped strandlines.

The domelike form is revealed by the isobases showing in meters the accumulated uplift during the past 7000 years (Fig. 79), the dating being based on the Swedish varve chronology of De Geer. As the isobases showing the present rate of uplift are drawn in terms of centimeters per century (= meters per 10,000 years), comparison of them with the isobases showing uplift in meters during the last 7000 years, which have much the same distribution, indicates that the present rate is only about two-thirds of the average rate during the past 7000 years.[15] This slowing down of the rate of uplift is inferred independently from other evidence

[13] Gutenberg 1941.

[14] If the current uplift from place to place is even only roughly proportional to the thickness of the former ice sheet, as it would seem that it should be, the form of the uplift suggests an ice sheet with a broad flat central area and with steep slopes at the extreme margins, especially the western margin, which should have received the greatest snowfall. (Cf. Chapter 15.)

[15] Gutenberg 1941, p. 739.

and is well established. Niskanen,[16] in part using estimates by earlier students, calculated that in Ångermanland on the western coast of the Gulf of Bothnia, where the ice sheet is known to have been thickest, the total crustal uplift prior to the year 6800 B.C. in the Swedish varve chronology amounted to about 270 meters, the uplift since 6800 B.C. amounted to 250 meters more, and the uplift still to be expected is 210 meters additional. This adds up to a grand total of 730 meters[17] of crustal subsidence under the weight of the ice in this district. From this value Niskanen calculated the corresponding thickness of the ice sheet to have been 2650 meters (= 10,433 feet). This figure compares favorably with estimates based on glacial-geologic evidence. These range from 2000 to 3000 meters.

The current updoming is causing all water bodies in the Baltic region to encroach upon their southeastern shores and to recede from their northwestern shores. The three connections between the Baltic and the North Sea (aside from the artificial Kiel and Göta canals) are the Öresund, the Great Belt, and the Little Belt. Their minimum depths are, respectively, 23 feet, 36 feet, and 36 feet. Thus uplift of 36 feet relative to sealevel would convert the Baltic into a lake or chain of lakes. Uplift now in progress will result, according to Niskanen,[18] in the emergence of the floor of the Gulf of Bothnia, so that 8000 years hence the northern part of the Gulf will consist of a small lake unconnected with the sea.

Updoming in North America

Updoming of the glaciated region has taken place in eastern North America, but, though a number of excellent studies of the problem have been made, the resulting synthesis is far less complete than the Fennoscandian synthesis. The difference is due in part to the much greater size of the North American area, in part to the scantier distribution of strandlines within the region studied thus far, and in part to the physical difficulties involved in examining in detail the vast uninhabited region of Canada where the ice sheet was thickest during the closing phases of the last deglaciation. We recognize the fact of postglacial updoming around the entire region vacated by the vast Laurentide Ice Sheet, but for the most part we know this fact only in a sketchy reconnaissance way. As yet we have more detailed data only from that sector which reaches from North Dakota to Newfoundland, not much more than a quarter of the

[16] Niskanen 1939.

[17] Gutenberg (1941, p. 762) gave 250, 250, and 200 meters, respectively, totaling 700 meters, of which 50 to 100 meters represent elastic deformation, the rest plastic.

[18] Niskanen 1939, p. 24.

whole arc of the former ice sheet. The immediate future offers splendid opportunities to continue this work through the abundant lacustrine strandlines of the Great Plains of Canada and the even more abundant marine deposits of the Canadian Arctic, the Hudson Bay region, and Labrador. Our discussion of North America, then, must be confined largely to the southern sector of the glaciated region.

Historical Review

As early as 1884 Upham recognized that the shorelines of the extinct Lake Agassiz are not horizontal, but he attributed this fact to the gravitative attraction of the ice sheet.[19] A year later Gilbert observed that strandlines in the Lake Ontario basin were deformed,[20] and soon afterward he rightly inferred that the deformation had resulted from removal of the glacier load. Before long Spencer inferred what he called the "focus" of uplift of the Great Lakes region to lie in the region southeast of Hudson Bay.[21]

The first attempt at broad regional synthesis was made by De Geer, who, after several years' study of the postglacial uplift recorded in Sweden, visited America in order to make comparisons. The map he published as a result shows contours on the imaginary surface of greatest uplift at each place throughout the sector from Newfoundland to Manitoba.[22] Deformation of the shorelines of Lake Bonneville had been shown by contours in Gilbert's classic monograph two years earlier, but De Geer called his contours *isobases,* a term that has been used for them ever since. De Geer's synthesis showed that the postglacial updoming of North America was closely analogous with that of Fennoscandia. A somewhat similar map, employing more data, though defective in some respects, was published many years later by Fairchild.[23]

We owe our knowledge of the warping of the classic Great Lakes region chiefly to the researches of Gilbert, Spencer, Leverett, Taylor, Coleman, J. W. Goldthwait, and Stanley. The data available up to 1915 were gathered into a classic summary by Taylor. This work, although amplified and modified in some ways, still stands as a monumental reference volume.[24] There is some information on the Hudson-Champlain depression,[25] though, for New England in general, data are

[19] Upham 1884.
[20] Gilbert 1885.
[21] Spencer 1889.
[22] De Geer 1892.
[23] Fairchild 1918, p. 202.
[24] Leverett and Taylor 1915, pp. 316–518. See also J. W. Goldthwait 1910*a*; 1910*b*.
[25] Woodworth 1905; Chapman 1937.

still scanty.[26] Data on eastern Canada have been gathered mainly by J. W. Goldthwait in numerous publications of the Geological Survey of Canada, of which the most extensive deals with Nova Scotia.[27] Data on Newfoundland are given by Daly and by Flint.[28] The existing information on the vast peninsula of Quebec has been summarized by H. C. Cooke.[29] Except for the Great Lakes region, knowledge of postglacial crustal warping in North America is in an undeveloped state, and much remains to be done before we can hope for a clear picture.

Inference from Warped Strandlines

As in Fennoscandia, we draw our inferences about the updoming chiefly from warped strandlines and from tide-gage records. From the warped strandlines, mainly in the Great Lakes region where they are best developed, but also to a lesser extent elsewhere in North America, can be drawn several important inferences:

1. The position of the outer limit of the deformed region is approximately determined, at least through the southern sector of the glaciated region.
2. In this same sector the outer limit of warping is shown to have shifted inward concentrically, and the rate of this shift *relative to the succession of Glacial Great Lakes* can be determined.
3. The rate of upwarping is shown to have been discontinuous in time. The discontinuities can be dated with respect to the succession of water bodies.

Let us consider these three general inferences in some detail.

OUTER LIMIT OF UPDOMING. The limits of the region domed up as the weight of the ice sheet diminished are determined approximately from the vertical positions of emerged strandlines of former glacial lakes and the sea. In the basins of Lake Erie and Lake Huron for example, the strandlines of Glacial Lakes Whittlesey and Maumee[30] are horizontal from their southern limits northward as far as Lake St. Clair and eastward as far as Ashtabula, Ohio. At these two localities they begin to rise, and they continue to rise northward to the points where each of them ended against the glacier margin. A line connecting these two localities

[26] Tarr and Woodworth 1903; J. W. Goldthwait 1925; Antevs 1928*b*; Hörner 1929; Flint 1933; I. B. Crosby and Lougee 1934; Perkins 1935; Hyyppä 1939; Jahns and Willard 1942.

[27] J. W. Goldthwait 1924.

[28] Daly 1921; Flint 1940*a*.

[29] H. C. Cooke 1930.

[30] The names and sequence of the glacial lakes referred to hereafter are given in Chapter 13.

is termed the *Whittlesey hinge line* (Fig. 81) because the upwarping of the Whittlesey (and Maumee) strandlines "hinged" along it. Figure 80 shows the ideal development of hinge lines.

Likewise the Lake Chicago strandlines that were formed at about the same time as Lakes Maumee and Whittlesey are horizontal from Chicago north to Milwaukee, Wisconsin, and Grand Haven, Michigan. From these places the strandlines rise northward until each comes to an

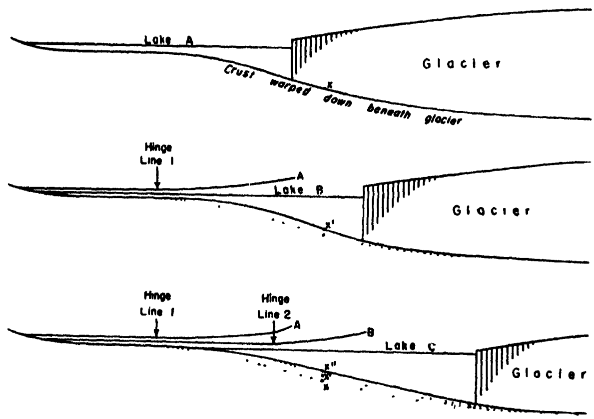


FIG. 80. Ideal profile showing progressive upwarping of the crust during deglaciation.

Upper profile. Glacial lake *A* has formed between the receding glacier margin and higher ground to the left.

Middle profile, later phase. Uplift has affected the area to the right of Hinge Line 1, bending the strandline of lake *A* to a higher position. Point *x* has been bent up to position *x'*. A new lake, *B,* has formed at a lower level.

Lower profile, still later. Renewed uplift has affected the area to the right of Hinge Line 2, bending the strandline of lake *B,* and warping the extreme right-hand end of *A* a little more. Point *x* has been bent up still more, to position *x"*. A third lake, *C,* has formed at a still lower level.

end. Accordingly it is inferred that the Whittlesey hinge line passes across Lake Michigan through these two places.

Although this evidence proves that there was no upwarping in the lakes region south of the Whittlesey hinge line after Lakes Maumee and Chicago had begun to form, it does not prove that some upwarping of that region did not take place in the long interval between the Wisconsin maximum and the development of the lakes. Accordingly the Whittlesey hinge line represents merely the outer limit of *measurable* crustal warping in the Great Lakes region. Traced westward, this outer limit (whether or not it coincides in time of origin with the Whittlesey hinge line) trends northwesterly, passing north of Glacial Lake Dakota, whose

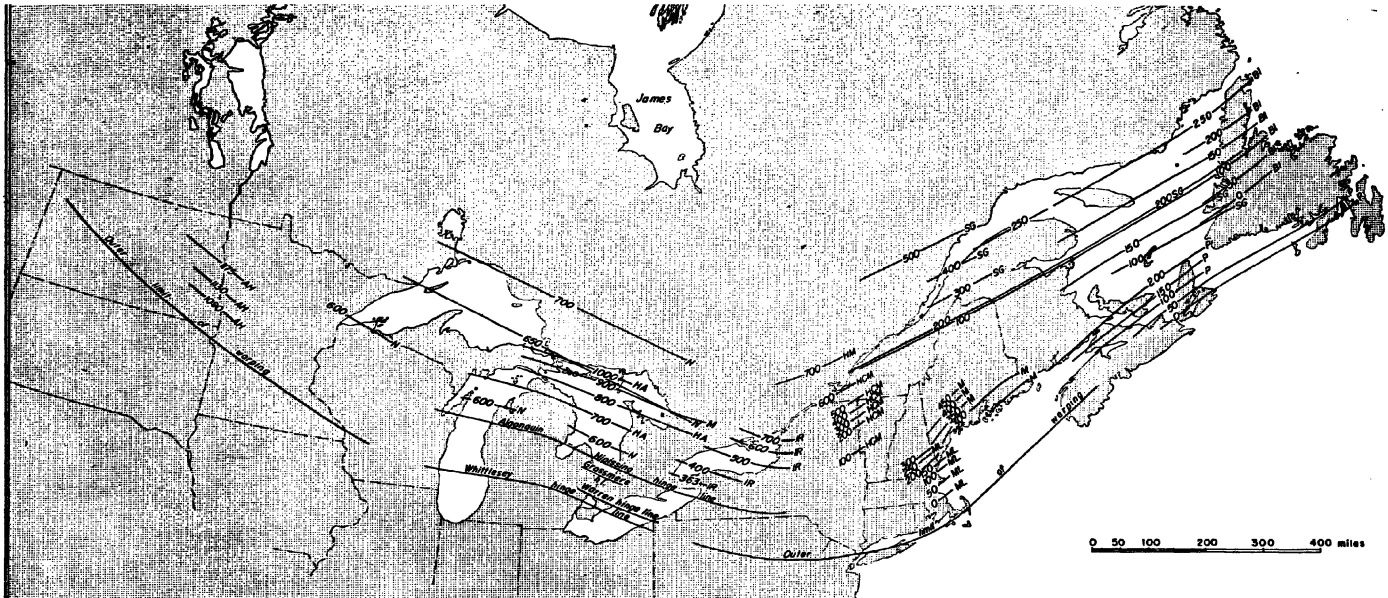


FIG. 81. Postglacial warping in the southern sector of the North American ice sheet. (Compiled from many sources.)

The curves are based on data of varying reliability and are only provisional, but they show clearly the form of the warping and the migration of the hinge line.

All the curves not otherwise labeled are isobases, and all figures are altitudes in feet above sealevel. Reading from left to right, the groups of isobases are: AH = Herman shoreline of Glacial Lake Agassiz. N = Nipissing Great Lakes shoreline. HA = highest shoreline of Glacial Lake Algonquin. IR = highest shoreline of Lake Iroquois. HCM = highest marine features in the Lake Champlain district. HM = highest marine strandline in Ottawa district. ML = "marine limit" in coastal Massachusetts-New Hampshire. M = highest marine features in coastal Maine. P = "Placentia group" of marine features. SG = "St. Georges group" of marine features. BI = Bay of Islands marine shoreline. Data from J. W. Goldthwait 1910*a*, 1910*b*, 1924, 1925; Leverett and Taylor 1915; Leverett 1929*a*; H. C. Cooke 1930; I. B. Crosby and Lougee 1934; Perkins 1935; Alcock 1935; Chapman 1937; Flint 1940*a*.

strandlines are said not to be warped, but south of Lake Agassiz, whose strandlines are warped even as far south as the lake outlet. On similar evidence the outer limit of warping is known to pass southwest of Glacial Lakes Souris and Regina. Although many strandlines are known to exist throughout the entire region between southern Manitoba and the Arctic Sea, we have no information whatever on their attitudes, and in consequence we are not yet able to specify anything about the postglacial warping of that vast region. We do know that much of the Pacific coast has been postglacially upwarped. On Vancouver Island postglacial marine deposits occur up to 400 feet above sealevel, and near Vancouver up to 650 feet.[31] In extreme northwestern British Columbia similar deposits occur up to 400 feet.[32] In southeastern Alaska[33] postglacial marine sediments occur at many places; locally they contain a cold-water molluscan fauna. The highest altitude thus far reported is 600 feet. In northwestern Alaska an authoritative summary[34] fails to report such features; presumably they do not exist. This presumption is consistent with the slight glaciation of northwestern Alaska, whereas the glaciation of southeastern Alaska and British Columbia was heavy.

Probable uplift of the glaciated Grand Coulee district in Washington has been inferred,[35] and, when the strandlines of the former great Lake Missoula in western Montana have been measured, probably they will prove to record upwarping. All these data, of course, are referable to the removal of the weight of the Cordilleran Glacier Complex. Although the Cordilleran glaciers were coalescent for a time with the Laurentide Ice Sheet along the eastern base of the Rocky Mountains, it is not likely that the movement was common, as a unit, to both glacier masses. Future discoveries will probably reveal that movements related to the two masses were quite distinct.

East of Lake Erie the data for fixing the outer limit of updoming are scanty, chiefly because strandlines are few. The northward rise of glacial-lake-bottom silts and clays (though conspicuous strandlines are absent) in the Hudson River and Connecticut River valleys indicates that warping affected the region at least as far south as Newburgh, New York, and Hartford, Connecticut. On the other hand there is no clear evidence in the record of Glacial Lake Passaic that warping affected the district southwest of New York City. The presence of fossil-bearing marine deposits overlying the Wisconsin drift in southeastern Massachusetts,[36]

[31] Johnston 1923, pp. 39–53.
[32] Hanson 1934, p. 181.
[33] Buddington and Chapin 1929, p. 276.
[34] P. S. Smith and Mertie 1930.
[35] Flint 1935*a*.
[36] Hyyppä 1939.

despite the eustatic rise of sealevel that has occurred since that district was deglaciated, shows that upwarping must have taken place there also.

From Massachusetts northward the same kind of evidence is common throughout the coastal region of New Hampshire and Maine. Warped marine strandlines occur in the eastern Maritime Provinces of Canada,[37] in southeastern Newfoundland,[38] and on the Coast of Labrador,[39] indicating that this entire region was affected. Emerged marine features are commonly reported from many parts of the Canadian Arctic. Their presence proves that uplift has occurred, but, because the attitudes of the strandlines have not yet been measured, the form of the uplift is not known, and it is not even certain (though it is surely very probable) that the uplift is of glacial origin.

INWARD MIGRATION OF HINGE LINE. The position of the Whittlesey hinge line, the southern limit of warping of the Whittlesey strandline, has been indicated. The hinge line of the Lake Warren plane lies northeast of the Whittlesey hinge line. The hinge line of the Grassmere and Elkton (Lundy) planes lies farther northeast, and the single hinge line common to both the Algonquin and the Nipissing lakes lies still farther in the same direction (Fig. 81). As these lakes succeeded each other in time in the order named, it is clear that between Whittlesey time and Nipissing time the outer limit of warping in the Great Lakes sector migrated inward (through a measured distance of about 75 miles). The tide-gage records, discussed in the next section, indicate that since Nipissing time the hinge line has migrated still farther north. It is probable that the same sort of migration took place, though not exactly contemporaneously, all the way around the center of the glaciated region; however, the facts as yet do not verify this assumption.

In contrast with the record of warping in the Great Lakes region, the record in Fennoscandia does not exhibit this hinge-line migration. There, as is apparent from comparison of Figs. 75, 76, and 77, the hinge lines of each of the successive water bodies seem to have occupied nearly the same position. The difference may lie in the fact that in Fennoscandia the water bodies lay nearer the center of the ice sheet than did the North American lakes. Therefore they did not form until a greater proportion of the maximum volume of the ice had wasted away, and hence a greater proportion of the total crustal adjustment had occurred, than had taken place in North America before the great glacial lakes appeared.

DISCONTINUITY OF UPLIFT IN THE GREAT LAKES REGION. The positions of the uplifted strandlines in the Great Lakes region indicate that in that

[37] J. W. Goldthwait 1924.
[38] Flint 1940*a*.
[39] Daly 1902.

region there were at least three times of active uplift separated by two times of crustal stability:

1. The first recorded uplift began early in the Mankato sub-age just before Lake Whittlesey ceased to exist. At that time nearly half the Great Lakes region had already been relieved of its load of ice, as shown by the northward extent of the Lake Whittlesey strandline. The movement continued intermittently throughout the lives of Lakes Warren and Lundy and into the time of Lake Algonquin I. This inference is based on the fact that the Whittlesey, Warren, Lundy, and earliest Algonquin strandlines are inclined at different angles and have different hinge lines.

2. Crustal stability began with Lake Algonquin II (with the Trent Valley outlet) and continued through Algonquin III (the time when three outlets existed simultaneously) into Algonquin IV (with the North Bay-Ottawa outlet). Stability is shown by the facts that the strandlines (including those of later Lake Vermont) referable to this time are nearly parallel, and that the highest Algonquin strandline and the strandline of the contemporaneous Lake Iroquois are strongly developed, implying a long pause at these levels.

3. The second and greatest uplift set in in the later part of the Algonquin IV phase, lifting the North Bay-Ottawa outlet so high that the overflow from the upper lake basins was again sent south to Lake Erie. This uplift is recorded by the strong northeast uplift of the nearly parallel Algonquin strandlines down through and including the Payette. The uptilts of all these strandlines are referable to a single hinge line. This movement continued, possibly with some interruption, almost to the beginning of the Nipissing Great Lakes phase.

4. The second time of crustal stability coincides essentially with the life of the Nipissing Great Lakes. This quiescence is shown by the unity and the mature development of the Nipissing strandline, which was evidently fashioned by waves and currents that operated at one level for a long period.

5. The third uplift began in post-Nipissing time, gently warping the Nipissing strandline up toward the northeast. It may have continued uninterruptedly into the modern movement revealed by tide-gage measurements.

It is highly probable, though not demonstrated, that the cause of discontinuity of uplift lies in discontinuity of deglaciation: that is, the times of uplift coincide with or follow times of shrinkage of the ice sheet, whereas episodes of quiescence coincide with or follow episodes of equilibrium in the glacier's regimen or even of expansion of the ice sheet. If this is true, then the basic cause of the alternation of movements and quiescences consists of climatic changes of small magnitude, and these

may have varied somewhat, perhaps greatly, from place to place along the vast peripheral arc of the ice sheet. In fact, when the Whittlesey and Algonquin hinge lines and the Algonquin and Nipissing isobases in the Great Lakes region are plotted on a single map (Fig. 81), it is apparent that they converge toward the east. This convergence suggests that, during the time represented by these features, deglaciation was taking place more rapidly in the western part of the region than in the eastern. This in turn may have resulted from meteorologic conditions, analogous to those of today, which nourished the waning ice sheet more effectively along the Mississippi Valley storm tracks than in the district farther northwest.

It is interesting to compare the two discontinuities recorded in the Great Lakes region with those recorded in the Baltic region and shown in Fig. 78. As the facts become better known such breaks may prove to furnish an important means of correlation, if they are the result of worldwide climatic fluctuations. But information at present is too scanty to justify such correlation.

Discontinuity of uplift elsewhere. Knowledge of the three movements of uplift in the Great Lakes region still awaits extension toward the west. Of the strandlines of lakes Regina, Souris, and Agassiz (Fig. 81) we can say that they are warped up toward the north and east, the highest ones the most steeply in the case of Agassiz. But the unraveling of the discontinuities awaits detailed measurements that have not been made.

Of New England and eastern Canada, likewise, little is known, because the strandlines from which the critical information must come have not been studied in enough detail. Probably this region will never be as well known as that of the Great Lakes, because its strandlines are less numerous, less extensive, and less well developed. At present it is known merely that warping occurred throughout New England and eastern Canada, and that the trend of the isobases, as far north as Newfoundland, is northeast (Fig. 81).

In Newfoundland there is clear evidence of discontinuity of uplift. At least three warped marine strandlines are present there.[40] The highest and oldest, apparently represented also in Nova Scotia, is warped steeply. The intermediate and the lowest strandlines, apparently well represented also in New Brunswick, are warped less steeply, the angles of slope of both being almost the same. The lowest, the Bay of Islands surface, is a wide bench cut by waves across bedrock and therefore records rather long stability. Tempting though it is to try to correlate these uplifts and quiescences with those clearly evident in the Great Lakes

[40] Flint 1940a.

region, we must remember that, if the movements are directly related to the shrinkage of the ice sheet, they were not necessarily synchronous through the 1500-mile distance that separates these two regions. Direct correlation by continuous strandline studies throughout the whole length of the St. Lawrence, the logical connecting path, will be necessary before connections can be made on a firm basis.

Inference from Tide-Gage Records

As early as 1896 study of the tide-gage records of the Great Lakes convinced Gilbert that this region is being warped up toward the north at the present time. Subsequent study of these records[41] has fully confirmed Gilbert's opinion and has extended it to include the Atlantic coast between New Jersey and Quebec. The average rate of tilt in the Great Lakes region is about 6 inches per 100 miles per century, sufficient to cause the lakes to rise against their southwestern shores while receding from their northeastern shores. In time the shoaling of some Canadian harbors due to this cause may become a serious problem. Gutenberg's map, showing curves indicating rate of uplift per century at 20-centimeter intervals, is here reproduced as Fig. 82. Comparison with Fig. 79 shows that the rate is very nearly the same as in Fennoscandia. The zero curve lies close to the position of the Algonquin hinge line, showing that the outer limit of updoming is still nearly the same as it was when Lake Algonquin was in existence.

Assuming that conditions in North America are analogous with those in Fennoscandia, Gutenberg calculated that the center of the dome, in the Hudson Bay region, should rise about 250 meters more before crustal recovery is complete.[42] This rise would reduce Hudson Bay to a tiny inlet of Hudson Strait in the extreme northeastern part of its present area. The Great Lakes region, being at a considerable distance from the center of the dome, should experience a smaller future uplift.

Updoming Elsewhere

The British Isles were depressed and then domed up independently of the movements in Fennoscandia, thus reflecting the presence of independent centers of radial outflow of ice in Britain. At least two postglacial marine shorelines have been identified, and their distribution clearly indicates that the thickest glacier ice lay over Scotland rather than over England (Fig. 83). The higher of the two shorelines now stands at

[41] Summarized and amplified by Gutenberg (1933; 1941, p. 739).
[42] Gutenberg 1941, p. 766.

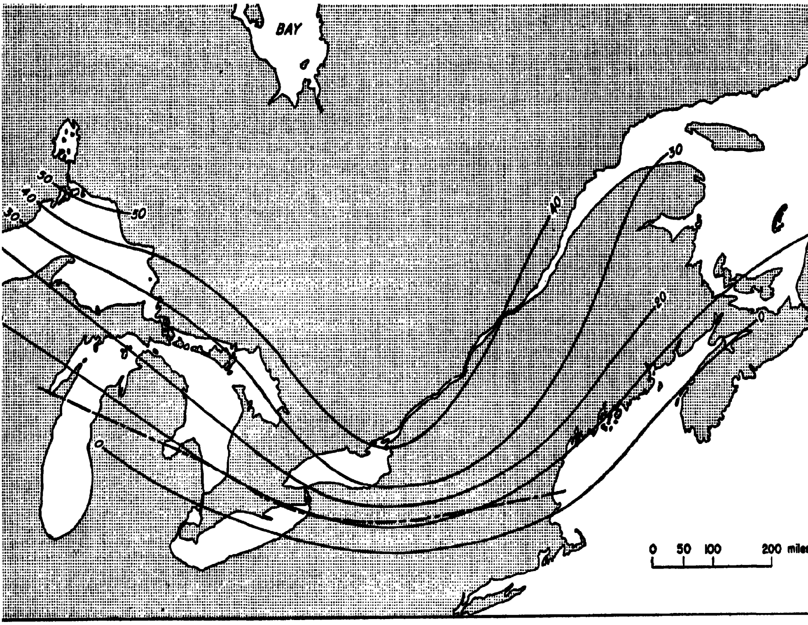


FIG. 82. Present rates of uplift in eastern North America.

The continuous curves indicate amount of uplift in centimeters per century. The broken line shows the position of the Algonquin hinge line, its eastward extension according to Gutenberg. (After Gutenberg 1941, p. 743.)

about 100 feet above sealevel in the district in which it has been domed up the most. The lower reaches a maximum of about 35 feet. The fashioning and subsequent updoming of the lower strandline took place very recently, as is shown by its relation to the rise of sealevel discussed elsewhere in this chapter.[43]

On the coast of Iceland delta terraces at the mouths of large streams occur at altitudes up to 250 feet above present sealevel. They are believed to have been deposited during the last deglaciation, after which they were warped up to their present positions as a result of lightening of the weight of glaciers on Iceland.

Emerged marine beaches are common on the west coast of West Spitsbergen. The highest stands at 430 feet. Uplifted postglacial marine features in Novaya Zemlya reach about 300 feet above present sealevel, indicating at least 300 feet of upwarping since the deglaciation of that region began. Data on Greenland are scanty. Uplift of at least 300 feet is clearly recorded in the Fiord region of East Greenland, and higher figures are reported from districts farther north. In the glaciated parts of the southern hemisphere, uplifts of at least 125 feet in Patagonia and 300 feet in Antarctica have been reported.

[43] Godwin 1943.

In summary, we find that those parts of the Earth's surface that have sustained or are sustaining large ice sheets subsided under the extra loads and gradually recovered as the loads were removed. The best-known regions of subsidence and recovery are the regions of the former large ice sheets of Europe and North America. In both regions recovery has been discontinuous; in both it is still in progress. And in both regions the deciphering of the crustal movement is complicated by the changes of sealevel now to be discussed.

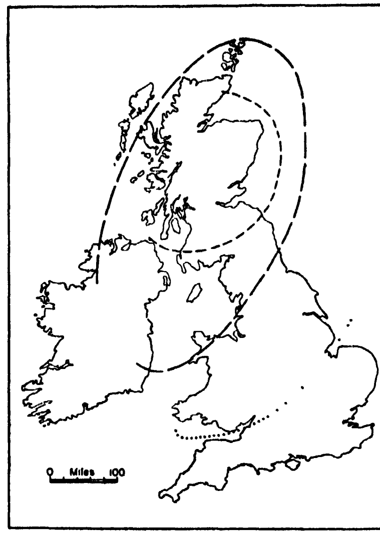


FIG. 83. Postglacial upwarping in the British Isles.

Short-dash line = limits of the "100-foot beach" formed during deglaciation. (W. B. Wright.)

Long-dash line = limits of the "25-foot beach" formed after deglaciation. (W. B. Wright.)

Dotted line = southern limit of the updomed region. (Zeuner.)

FLUCTUATION OF SEALEVEL

Maclaren, who first deduced that extensive glaciation would reduce the level of the sea, estimated that the lowering must have been of the order of 350 to 700 feet.[44] Since the appearance of Maclaren's discussion there has been general agreement that growth and decay of glaciers have been accompanied by falling and rising of sealevel, but as to the amount of sealevel fluctuation opinions have differed widely. These large differences have arisen from the fact that throughout most of the hundred years since Maclaren's time the facts available have not been adequate to confine within narrow limits the many calculations, speculations, and guesses that have been made. The figures on record range from "less than 1 millimeter" to many thousands of feet. It would doubtless have been a great satisfaction to Maclaren if he could have known that at the end of a century his minimum figure is now thought to be of the right order of magnitude, for it is now believed that at the time of the most extensive glaciation the sealevel was reduced 350 to 400 feet, that in the Fourth Glacial age it was reduced 230 to 330 feet, and that, if all existing

[44] Maclaren 1842, p. 365.

glaciers were to melt, the sealevel would be raised 65 to 165 feet above its present position.[45]

EVIDENCE FROM TIDE-GAGE RECORDS

The fact that the sealevel is now rising upon the lands is shown by the records of tide gages, some of which date back to the middle of the nineteenth century. Each gage records the change of sealevel relative to the land at one place, but it does not tell the cause of the change. In the central parts of the recently glaciated regions the tide gages record a relative lowering of sealevel, which, as has been shown, is the result of doming-up of the crust in these regions. Elsewhere, in some parts of the Mediterranean, for example, relative lowering of sealevel is similarly recorded. Most such places are in regions of known crustal instability, and the records are believed to indicate local upward movement of the crust. The great majority of records, however, aside from these two special cases, combine statistically to show a rise of sealevel, relative to the land, throughout the last few decades, at a rate of about 2.5 inches per century.[46] This figure gives a picture of the slow worldwide rise of sealevel that is occurring independently of local warping of the crust. Probably a small part of this rise of sealevel is the result of isostatic upwarping of glaciated regions of shallow sea, such as Hudson Bay and the Gulf of Bothnia, causing them to spill their waters, as it were, into the main ocean body. Another part may be due to the current addition of sediment to the sea, displacing water and therefore lifting the sealevel. But the bulk of the rise is attributable to the melting of glaciers.[47] This rate of melting, however, has been considerably greater during the last few decades than it was during the preceding century, and from our record of climatic changes during historical time, very imperfect though it is, we can reasonably conclude that the level of the sea has risen at a variable, perhaps an extremely variable, rate.[48]

[45] These figures are obtained by calculating the volumes of water that are believed to have been locked up in the glaciers at these times. They do not take account of the isostatic depression or elevation of the sea floor that would accompany the addition or subtraction of layers of water of these thicknesses (see A. C. Lawson 1940). If this factor is considered, the figures given above must be somewhat reduced. Still other factors are involved (see Daly 1934, p. 48).

[46] Gutenberg 1941, p. 730; Marmer 1943.

[47] Cf. Thorarinsson 1940.

[48] Interesting evidence on this matter is presented by Godwin (1943).

Deductions from Calculated Volumes of Glacier Ice

Existing Glaciers

If the volume of all the glacier ice that rests on the land today were known, we could calculate very roughly the amount of rise of sealevel that would take place if all this ice were to melt. Floating ice is, of course, left out of consideration, because its melting has no effect on the level of the sea. The area of existing glacier ice (given in Table 1) is fairly well known, but its thickness, especially in the Greenland and Antarctic ice sheets, as yet can only be guessed. Because of the very uncertain factor of thickness, estimates of existing volumes of ice are unreliable and will probably remain so until reliable geophysical traverses are run across the great ice sheets in several directions. All the volume estimates made thus far are of the same order of magnitude because the fundamental assumptions as to the thicknesses of the great ice sheets have been comparable.

In the calculation of thickness it must be remembered that the crust subsided beneath the weight of each large ice sheet. In theory this factor would increase the probable thickness of the ice and hence the reduction of sealevel. However, as evidence of thickness is extremely scanty at best, the factor has little practical value at present.

Table 18. Estimated Volume of Existing Glacier Ice According to Antevs*

	Area (*km²*)	*Average thickness* (*meters*)	*Volume* (*km³*)
1 Antarctic Ice Sheet	13,500,000	1000 to 1525	13,500,000 to 20,587,500
2. Greenland Ice Sheet	1,833,900	1400	2,567,460
3. All other glaciers	256,000	Small	Neglected
Total			16,067,460 to 23,154,960
Corresponding rise of sealevel†			40 to 60 meters

* Antevs 1929*a*, p 43.

† Isostatic readjustment is neglected. Area of the sea is taken as 361,059,200 km². Specific gravity of ice is taken as 0 91.

Another uncertain factor, neglected in some calculations, is the subsidence of the sea floor under the weight of water added to the sea. Calculations made by Lawson[49] are based on the assumption that the amount of subsidence will equal about one-third the thickness of the layer of water added. Daly's calculations[50] imply an adjustment of somewhat smaller value.

[49] A. C. Lawson 1940.

[50] Daly 1934.

TABLE 19. ESTIMATED VOLUME OF EXISTING GLACIER ICE ACCORDING TO RAMSAY*

	Area (km^2)	Average thickness (meters)	Volume (km^3)
1 Glaciers in the temperate and warm zones	70,000	150	10,000
2. Smaller glaciers in the polar regions	186,000	300	370,000
3. Greenland Ice Sheet	1,870,000	1400	2,620,000
4. Antarctic Ice Sheet	13,000,000	1500	19,500,000
Total			22,500,000
Corresponding rise of sealevel			55 meters

* Wilhelm Ramsay 1930, p 15.

Shortly after the appearance of Ramsay's estimate, Dubois calculated that the melting of existing glaciers would raise the sealevel 38 to 58 meters.[51]

TABLE 20. ESTIMATED VOLUME OF EXISTING GLACIER ICE ACCORDING TO DALY*

	Area (km^2)	Average thickness (meters)	Volume (km^3)
1. Greenland Ice Sheet	1,834,000	1000	1,834,000
2. Antarctic Ice Sheet	13,500,000	1400	18,900,000
3. All other glaciers	500,000	300	150,000
Total	15,834,000		20,884,000
Corresponding rise of sealevel			50 meters†

* Daly 1934, p 12 Floating glacier ice is neglected.
† Adjustment reduces this figure to "less than 40 meters."

Thorarinsson[52] recalculated the *areas* of existing glaciers, as shown in Table 21.

Thorarinsson's areal figures are undoubtedly more accurate than those made earlier. The thickness figures here applied to them are very conservative, in line with the growing belief that thicknesses in Greenland and the Antarctic Continent are much less than was formerly thought. The resulting figure of 24 meters is as likely to lie near the truth as the earlier, larger estimates. Not until more is learned about the thicknesses of the existing ice sheets will the approximate truth be known.

On the basis of these various estimates it appears that melting of all the existing glacier ice would raise the sealevel 24 to 60 meters, less an

[51] Georges Dubois 1931.
[52] Thorarinsson 1940.

TABLE 21 ESTIMATED AREA OF EXISTING GLACIER ICE ACCORDING TO THORARINSSON

(Estimated thicknesses and computed volumes are added to the figures modified from Thorarinsson, cited in Table 1.)

	Area (km^2)	*Average thickness* (*meters*)	*Volume* (km^3)
1. Continental Eurasia	123,320	200	24,418
2 Pacific Islands	1,015	100	101
3. Continental Americas	105,000	200	21,000
4. South Polar Region	13,003,000	600	7,801,800
5. North Polar Region	1,867,700	900	1,680,930
Total	15,100,035		9,528,249
Corresponding rise of sealevel (less isostatic adjustment)			24 meters

allowance (say 5 to 10 meters) for isostatic adjustment. Thus actually the sea would stand 20 to 50 meters (66 to 165 feet) higher against the lands than it does today.

It is important that the accuracy of the basic data be improved because the result represents the maximum height (neglecting other factors) that sealevel could have attained during any interglacial age and thus affects not only the problem of abandoned strandlines but also the study of migrations of animals and plants during interglacial and preglacial times.

Former Glaciers

It is even more important to attempt to estimate the amount of lowering of sealevel that was brought about by the growth of the glaciers to their maximum former extents. But this estimate is also more difficult, both because we can not closely determine the factor of thickness of the former ice over the greater part of the glaciated regions and because we have to assume without proof that during any glacial age the glaciers reached their maxima at the same time. Many estimates have been made, but the earlier estimates are of only historical interest because of inaccuracy of the volume data on which they are based.

The area of a former ice sheet can be determined from geologic evidence. Mean thickness can not be determined accurately; in fact it can only be roughly approximated. The ice sheets referable to the several glacial ages have to be clearly distinguished from one another. Because of inaccurate data and inadequate correlations this means that for the purpose of volume determination only two time groups of glaciers can be recognized as yet: those referable to the Fourth Glacial age, and those referable to the maximum areal extent of all glaciers. It has to be assumed

(though not without confirmatory evidence) that the glaciers reached this maximum more or less contemporaneously in all parts of the world. As for the other glacial ages, the evidence from the areal relations of the drift sheets in both North America and Europe implies that at the times corresponding to them the areas of the ice sheets were, on the average, intermediate between those referable to the maximum glaciation

TABLE 22 ESTIMATED VOLUME OF GLACIER ICE AT THE MAXIMUM OF THE FOURTH GLACIAL AGE, COMPILED FROM ANTEVS*

(Note: All figures indicate *excess* over volumes of existing glaciers Question marks indicate especially doubtful estimates)

	Area (km^2)	*Average thickness* (*meters*)	*Volume* (km^3)	
North American Ice Sheet (East of the Cordillera)	9,000,000	2700	24,300,000	
Cordilleran Glacier Complex	2,500,000	1100	2,750,000	
Total North America				27,050,000
European Ice Sheet† (East of Scandinavian Mts.)	2,200,000	2000	4,400,000	
(West of Scandinavian Mts.)	1,100,000	500	550,000	
Total Europe				4,950,000
Eurasian glaciers†	3,500,000	100?		350,000?
Greenland Ice Sheet				400,000
Antarctic Ice Sheet	13,500,000+	300		4,050,000+
South American glaciers				50,000
Grand total				36,850,000
Corresponding lowering of sealevel				90 meters§

* Antevs 1928*a*, pp. 74–82.

† Gerasimov and Markov (1939, p. 445) assign somewhat different areas to the glaciers in Europe and Asia and consider that the glaciers in these two regions were not entirely contemporaneous.

§ Less the effect of isostatic adjustment.

and those referable to the latest glaciation. Hence the corresponding sealevels should be intermediate between those arrived at for the Fourth Glacial age and for the time of maximum extent.

Tables 22 and 23 give estimates of the volume of glacier ice that existed during the Fourth Glacial age (the Wisconsin age in North America) according to Antevs and according to Daly. The two sets of figures are similar because in both sets the general sources of data were the same. Table 24 gives the corresponding figures for the time when all the glaciers reached their Pleistocene maxima. It seems likely that the

TABLE 23. ESTIMATED VOLUME OF GLACIER ICE AT THE MAXIMUM OF THE FOURTH GLACIAL AGE, ACCORDING TO DALY*

	Area (km^2)	*Average thickness* (*meters*)	*Volume* (km^3)
North American Ice Sheet (East of the Cordillera)	9,000,000	2400	21,600,000
Cordilleran glacier complex	2,500,000	800	2,000,000
European Ice Sheet	3,500,000	1600	5,600,000
East Asiatic glaciers	3,000,000	100	300,000
South American glaciers			50,000
Others			100,000
Greenland Ice Sheet (excess over present volume)	2,000,000	300	600,000
Antarctic Ice Sheet (excess over present volume)	13,500,000	300	4,050,000
Grand total	33,500,000		34,300,000
Corresponding lowering of sealevel			85 meters†

* Daly 1934, p. 46.

† Assuming that the glacial maxima were everywhere contemporaneous. Daly reduced the figure to 75 meters to allow for noncontemporaneity, less the effect of isostatic adjustment.

TABLE 24. ESTIMATED VOLUME OF GLACIER ICE AT THE MAXIMUM OF PLEISTOCENE GLACIATION, ACCORDING TO DALY*

	Volume (km^3)
North American glaciers	27,000,000
European glaciers	7,000,000
Siberian glaciers	1,000,000
All others	7,000,000
Total	42,000,000
Corresponding lowering of sealevel	105 meters†

* Daly 1934, p. 48

† Reduced to 90 meters to allow for noncontemporaneity. Allowance for isostatic adjustment would reduce the figure still further.

deduced lowering of sealevel at the maximum of the last glacial age, 70 to 90 meters less an allowance for isostatic adjustment, at least approaches the right order of magnitude. The assumption that the glaciers reached their maximum extents contemporaneously, although manifestly not true in detail, is believed to be true in a broad sense. At least the probable error resulting from this assumption is likely to be smaller than the probable error resulting from the figures adopted for ice thickness.

TABLE 25. ESTIMATED VOLUMES OF GLACIER ICE AT MAXIMUM PLEISTOCENE EXTENT AND AT MAXIMUM OF THE FOURTH GLACIAL AGE

	At Maximum Pleistocene Extent				At Maximum Fourth Glacial Age			
	Areas Covered by Glaciers		Estimated Average Thickness of Ice (meters)	Computed Volume of Ice (cubic kilometers)	Areas Covered by Glaciers		Estimated Average Thickness of Ice (meters)	Computed Volume of Ice (cubic kilometers)
	square miles	square kilometers			square miles	square kilometers		
Laurentide Ice Sheet	5,071,000	13,133,890	2000	26,267,780	4,840,000	12,535,600	2000	25,071,200
Cordilleran glacier ice (coalescent)	957,000	2,478,000	900	2,230,200	875,000	2,266,250	900	2,039,625
Cordilleran (separate areas, including Mexico and Hawaii)	40,000	103,600	220	22,792	35,000	90,650	200	18,130
Greenland Ice Sheet	935,000	2,421,650	700	1,695,155	835,000	2,162,650	600	1,297,590
Total North America	7,003,000	18,137,140		30,215,927	6,585,000	17,055,150		28,426,545
Scandinavian Ice Sheet	2,145,000	5,555,550	1750	9,722,212	1,650,000	4,273,500	1750	7,478,625
Glaciers of British origin	176,000	455,840	1750	797,720	143,000	370,370	1750	648,147
The Faeroes	5,000	12,950	300	3,885	4,000	10,260	250	2,565
Iceland and Jan Mayen Island	55,000	142,450	700	99,715	45,000	116,550	550	64,102
Spitsbergen	77,000	199,430	600	119,658	60,000	155,400	500	77,700
Separate areas in continental Europe	15,000	38,850	300	11,655	13,000	33,670	250	8,417
Total Europe	2,473,000	6,405,070		10,754,845	1,915,000	4,959,750		8,279,556
Siberian Ice Sheet	1,628,000	4,216,520	450	1,897,434	836,000	2,165,240	350	757,834
Central Siberian Plateau	7,000	18,130	150	2,719				
Franz Josef Land	29,000	75,110	300	22,533	20,000	51,800	250	12,950
New Siberian Islands and Wrangell Island	51,500	133,385	150	20,007	35,000	90,650	130	11,784
Northeastern Siberia (coalescent glaciers)	440,000	1,139,600	200	227,920	360,000	932,400	175	163,170
Koryak Mountains	44,000	113,960	200	22,792	38,000	98,420	180	17,716
Kamchatka Peninsula	22,000	56,980	150	14,245	19,000	49,210	135	6,643
Transbaykal highlands (coalescent glaciers)	66,000	170,940	150	25,641	54,000	139,860	130	18,182
Other separate areas in Siberia	10,000	25,900	150	3,885	8,000	20,720	130	2,694
Altai highlands	176,000	455,840	100	45,584	125,000	323,750	80	25,900
Coalescent glaciers in central Asia, including Himalaya Mountains	440,000	1,139,600	150	170,940	335,000	867,650	130	112,794

Other separate areas in central and eastern Asia	60,000	155,400	100	15,540	42,000	108,780	80	8,702
Caucasus Mountains and separate areas in Asia Minor	5,000	12,950	120	1,554	4,250	11,007	100	1,101
Total Asia	2,978,500	7,714,315		2,470,794	1,876,250	4,859,487		1,139,470
Total *Northern Hemisphere*	12,454,500	32,257,155		43,441,566	10,376,250	26,874,387		37,845,571
Coalescent glaciers in southern South America	297,000	769,230	300	230,769	265,000	687,350	270	85,584
Separate areas in South America*	75,000	194,250	200	38,850	58,000	150,220	175	26,288
Total South America	372,000	963,480		269,619	323,000	837,570		111,872
Africa*	200	518	100	52	180	466	80	37
New Zealand, Tasmania, Australia, and New Guinea	25,700	66,563	180	11,981	22,500	58,275	160	9,324
Antarctic Continent	5,511,000	14,273,490	900	12,846,141	5,000,000	12,950,000	900	11,655,000
Total *Southern Hemisphere*	5,908,900	15,304,051		13,127,793	5,345,680	13,846,311		11,776,233
Grand Total	18,363,400	47,561,206		56,569,359	15,721,930	40,720,698		49,621,804
Less estimated volume of existing land ice				9,528,249				9,528,249
Excess of former ice over existing ice				47,041,110				40,093,555
Volume of water equivalent to excess ice (density of ice taken as 0 92)				43,277,720				36,886,070
Corresponding position of sealevel below present sealevel (less reduction for isostatic adjustment) (area of sea taken as 361,059,000 sq km.)				120 meters				102 meters
Proportion of present land area of world covered by ice (considering present land area as 150,274,572 sq. km)				32%				27%

* Some areas lie in the northern hemisphere but are too small to affect the total significantly.

Because we have to assume that these maxima were contemporaneous, and because the geologic evidence of the maximum glaciation is much less distinct than the evidence of the latest glaciation, the margin of error is probably greater here than in the calculations shown in Tables 22 and 23. Higher figures have been arrived at by others, notably 131 meters by Dubois[53] and 276 meters by Ramsay,[54] who assumed a thickness for the North American ice, and an extent and thickness for the Eurasian ice, that are almost certainly much too great.

Since the estimates in the foregoing tables were made, the extent of former glaciers, especially in Siberia, has become better known, and opinion has shifted toward the view that the existing ice sheets are thinner than was believed earlier. Table 25 takes these factors into account. Areas have been determined planimetrically.

In general Table 25 shows larger glaciated areas but thinner glaciers than those shown in Tables 22, 23, and 24. The resulting volumes of ice in both the Fourth Glacial age and the Pleistocene maximum are more than in the earlier estimates. Table 25 is not necessarily more accurate than Tables 22 and 23. All that can be claimed for it is that it is more nearly in accord with the facts now known.

It must be emphasized that the figures representing ice thicknesses are no more than thoughtful guesses. Minimum thicknesses are recorded at a few points by geologic evidence, but over most of the vast lowland areas covered by the former great ice sheets no means have been found for fixing ice thicknesses.

Unfortunately, however, the deduced amount of corresponding lowering of sealevel depends on the thickness of the ice sheets. Since thickness figures are highly uncertain, the tables give the appearance of greater certainty as to the deduced fluctuations of sealevel than is justified by the facts. A slight difference in the assumed thickness of a big ice body such as the Laurentide Ice Sheet makes a very large difference in the resulting contemporary level of the sea. For example the combined area of the Antarctic, Greenland, Laurentide, and Scandinavian ice sheets at the maximum of the last glacial age (exclusive of the Cordilleran, Siberian, and all other glaciers) was nearly 30,000,000 square kilometers, while the area of the sea is about 361,059,000 square kilometers. Considering the density of the ice as 0.92 and comparing these figures, we find that, for every meter added to the thickness of these ice sheets alone, the sealevel must have been lowered roughly 7.6 centimeters.[55] Since we

[53] Georges Dubois 1931, p. 657.

[54] Wilhelm Ramsay 1930, p. 49. Farrington (1945) reached a very high figure by inference from stratigraphic features, as outlined elsewhere in this chapter.

[55] Less deduction for the effect of isostatic adjustment.

do not yet know within some hundreds of meters how thick these ice sheets were, we are clearly not yet able to deduce glacial-age sealevels with great accuracy.

In summary, we find that deduction by various authorities from the estimated volumes of existing and former glacier ice supplies us with these figures:

Rise of sealevel if all existing glacier ice were melted, as during one or more interglacial ages	24– 60 meters[56]
Lowering of sealevel during the Fourth Glacial age	70–102 meters[56]
Lowering of sealevel during the maximum Pleistocene extent of glaciers	105–120 meters[56]

Further, these figures are subject to a large error which at present is not calculable.

Whatever their real value, these figures will have more significance if we compare them with the direct geologic record of Pleistocene fluctuation of sealevel.

Inference from Geologic Data

Fluctuation of sealevel during the Pleistocene epoch is abundantly demonstrated by geologic evidence, but the evidence is of many kinds and must be used with caution, and the actual amount of fluctuation can as yet be inferred only very imperfectly from the geologic record. Along many coasts the presence of marine features above sealevel and subaerial features below sealevel proves that emergence and submergence have occurred, but it is a quite different thing to demonstrate that the emergence and submergence resulted from fluctuation of the sealevel against a stable land rather than from crustal movement with stable sealevel, or even from a combination of the two activities. Because of the danger of confusion we must eliminate from consideration all coasts in regions now or formerly glaciated, for they are likely to have been affected by warping. We must eliminate also coasts in nonglaciated regions of crustal instability such as California and parts of the Mediterranean. In certain other coastal regions we have determined the presence not merely of marine deposits (which do not closely mark the position of sea*level*) but of strandlines, which can have been fashioned only at sealevel, and which are approximately horizontal throughout a distance great enough to show that the change of level resulted from fluctuation of sealevel rather than from local movement of the crust. This applies with equal force to features both above and below present sealevel.

But even here there is no certainty that the sealevel fluctuations resulted

[56] Less an uncalculated amount attributable to isostatic adjustment.

from glaciation rather than from crustal movements beneath some sea floor, that shifted the water and changed its level somewhat. All we can say at present is that the observed fluctuations of sealevel *seem* to fit into the pattern of the great fluctuations of the glaciers and promise closer comparison as future study enlarges our knowledge.

We have now to see what can be inferred from emerged marine deposits and from strandlines not now at sealevel. We shall consider the evidence of former higher sealevels first.

Evidence of Higher Sealevels

Low, stable nonglaciated coasts such as the Atlantic coast of North America south of latitude 40°, the coast of Brazil, and parts of the coast of China are promising fields for the study of emerged strandlines. The chief difficulty these coasts are likely to present is the presence of extensive deposits made continuously by large streams during the whole time that the sealevel was shifting from a higher to a lower position. Such deposits destroy strandlines as they are made and thus leave no topographic record of former higher positions of sealevel. The record would be present in deposits below the present surface, but on broad low coastal plains exposures more than a few feet in depth are very scarce. This is apparently the general condition along at least the western part of the Gulf coast of the United States, where a coalescent row of deltas, extending up to considerable heights above present sealevel, effectively masks evidence of the higher Pleistocene sealevels that are believed to have existed here.[57]

ATLANTIC COAST OF SOUTHERN UNITED STATES. On the Atlantic coast the evidence is better, although much work will have to be done before the whole sequence of events can be reconstructed. North of the drift border the emerged marine features, including both strandlines and fossil-bearing blankets of marine sediments, clearly attained their positions partly because of postglacial updoming, and they can not therefore be taken as evidence of higher positions of the sealevel itself. South of the drift border, however, widespread marine sediments and marine strandlines occur from New Jersey to Florida and continue westward on to the Gulf coast.[58]

The lowest distinct strandline (Fig. 84), named the Suffolk scarp from the town of Suffolk, Virginia, is a conspicuous wave-cut cliff up to 60 feet in height; its toe stands at 20 to 30 feet above sealevel. It is nearly

[57] This matter is discussed further in Chapter 14.

[58] These are described in some detail in Flint 1940*b*, which also gives an extensive list of references on this subject.

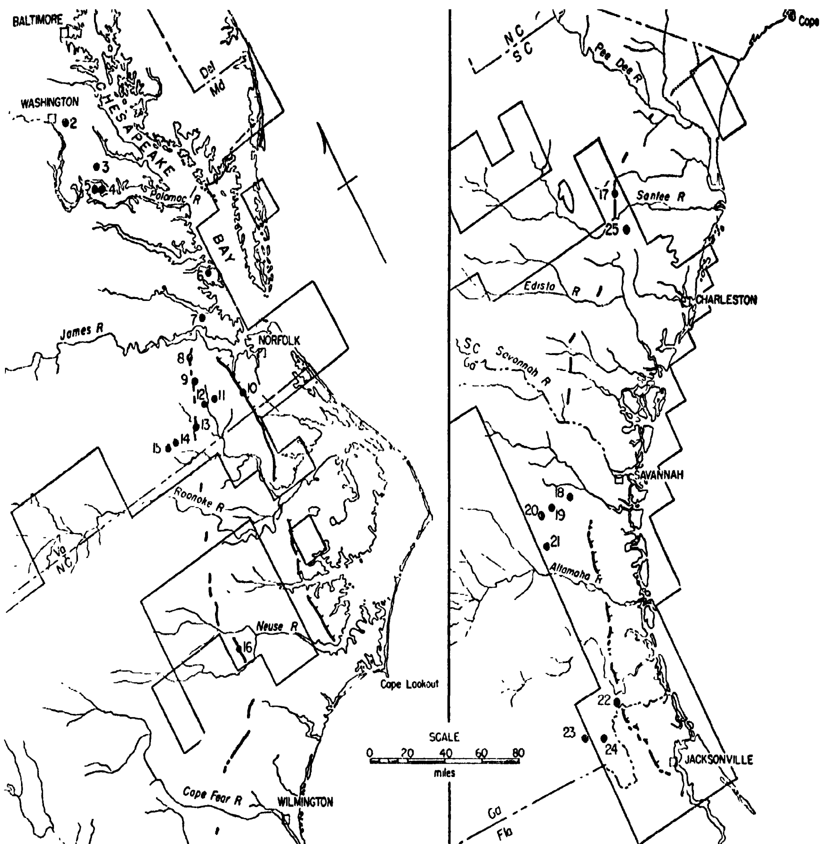


FIG. 84. Occurrence of the Suffolk scarp (hatched line; not mapped in South Carolina) and the Surry scarp (plain line) in coastal southeastern United States. Dotted line shows conspicuous bar at the altitude of the Surry scarp. Numbered points are sediment-sampling localities. (Flint 1940*b*.)

continuous throughout southern Virginia and North Carolina, and again through Georgia and much of Florida. Throughout most of South Carolina it is not present; in its place are numerous massive bars such as those that prevent cliffing by the waves at many places along the shoreline today. The Suffolk scarp is present in places along the coast as far north as New Jersey. Its maximum mapped extent (including gaps) is thus about 800 miles. Between the scarp and the present shore is a plain, in some places 50 miles wide, thinly covered with fine sediment. The deposits include a Pleistocene marine fauna that implies relatively warm water.[59] In New Jersey this sedimentary body has been called the Cape May formation. Farther south it has been called the Pamlico formation.

Ten to fifty miles back of the Suffolk scarp stands another, fainter and clearly older, wave-cut feature (Fig. 84) called the Surry scarp from the

[59] Richards 1936.

town of Surry, Virginia. Less prominent and less continuous than the Suffolk scarp, it is nevertheless sufficiently distinct to leave no doubt as to its origin. Its toe stands at 90 to 100 feet throughout its mapped extent, from the James River at the north to the Savannah River at the south, a distance of 375 miles. Near major streams the scarp is repeatedly interrupted by deltas and related deposits, and the plain to seaward of it is marked by scores of bars. The continuity and uniform altitudes of the Suffolk and Surry scarps, together with the sediments related to them, indicate that they originated when the sealevel stood higher than now and that their emergence has resulted from fluctuation of the sealevel against a stable land. Accordingly we can reasonably infer that these shores date from interglacial ages when glacier ice was less abundant on the lands than it is today. The Suffolk (Pamlico-Cape May) marine deposits overlie a zone containing cypress stumps in the position of growth.[60] This relationship shows that the Suffolk sealevel was preceded by a sealevel lower than the present one, and thus records a glacial age. We can not yet correlate these sealevels with the glacial succession. It is perhaps significant, however, that the altitude of the Surry sealevel, about 100 feet, is not much more than the altitude that would be expected if all existing glaciers were to melt away. Hence we may speculate that the Surry scarp dates from an extensive interglacial age in which glaciers, even in the polar regions, largely or wholly disappeared. The Yarmouth interglacial immediately suggests itself, but as yet this is hardly more than a speculation.

The warm-water fauna of the Suffolk (Pamlico-Cape May) deposits and the altitude of the Suffolk strandline together indicate that the Suffolk sea was interglacial; yet this sealevel must date from a time when a fair amount of glacier ice, though less than now, persisted in the polar regions. That time may be the Sangamon Interglacial. It could hardly be later than this, because, during the times intervening between the various Wisconsin glacial sub-ages, there appears to have been more rather than less ice on the lands than now.

Although the Surry scarp appears to be the highest record of Pleistocene marine overlap in Virginia and the Carolinas, there are marine bars at higher altitudes in southern Georgia and Florida. The highest and most conspicuous of them is Trail Ridge, whose crest slopes northward from 240 feet in north-central Florida to 160 feet in southeastern Georgia. The slope of Trail Ridge, which should have been more nearly horizontal when it was built, suggests that crustal warping has occurred here, probably at some pre-Surry time.

Not uncommonly the surface features in the region of marine overlap

[60] W. C. Mansfield 1928, p. 134.

have been described as *terraces*. Actually, however, the features consist entirely of marine plains, marine bars, wave-cut scarps, and, along the rivers, fluvial plains. To be sure, the area lying between the toe of the Surry scarp and the toe of the Suffolk scarp has a terracelike form, but its width is so great compared with its height that this form becomes conspicuous only when represented by a profile with the vertical dimension exaggerated. For this reason it seems best not to describe any groups of these features as *terraces*.

The surfaces of the interglacial seas must have stood, at least briefly, at many levels other than those recorded by the Surry and Suffolk scarps. Shorelines at higher, lower, and intermediate levels have been assumed,[61] and evidence of some of them is present in places. However, none has yet been shown to be as well marked or persistent as the two features named above.

The Surry scarp and the marine deposits related to it do not seem to be present north of the James River. Throughout the 280-mile distance from the James to the drift border the deposits and the surface forms that lie higher than the Suffolk-Cape May features are of fluvial origin. From this fact it may be inferred that while the sea stood at the toe of the Surry scarp the Potomac, Susquehanna, and Delaware river systems were building alluvium into the sea farther north. At that time the shoreline probably lay east of the present Chesapeake Bay and Delaware Bay shoreline. With later emergence, the alluvium was dissected and reworked by these same rivers, and it is represented today by the Beacon Hill, Bridgeton, and Pensauken sediments in New Jersey, and the Sunderland, Wicomico, and Talbot sediments in Maryland and Delaware.

On the Caloosahatchie River in south-central Florida is exposed at altitude less than 25 feet a section of alternating thin marine limestones and fresh-water marls separated by pronounced unconformities, the whole section being Pleistocene. The section is interpreted as recording four marine invasions and five lowerings of sealevel.[62] The relations are somewhat like those on Bermuda, described below.

BERMUDA. Interglacial sealevels higher than now are recorded on the island of Bermuda by two kinds of evidence: emerged wave-cut benches and Pleistocene marine limestones. A wave-cut bench at an altitude of 25 feet occurs at the two ends of the island.[63] It may be the Bermuda representative of the Suffolk sealevel. Two Pleistocene marine limestones, the Belmont and Devonshire limestones, form part of the Bermuda stratigraphic column. Both lie above present sealevel. From their strati-

[61] C. W. Cooke 1939, p. 33.
[62] Parker and Cooke 1944, p. 94.
[63] Sayles 1931, p. 446.

graphic relations Sayles believed their age to be interglacial and at least as old as Yarmouth. Unless (as has been shown to be very unlikely) the crust has risen here, these limestones prove a sealevel formerly higher than now.

Along other segments of the North American coasts, evidence of Pleistocene sealevels higher than now is so intricately associated with evidence of crustal movements that clear inferences as to fluctuations of sealevel can not be drawn safely. An exception may be the northwest coast of Alaska, a region that was not glaciated, where Pleistocene fossil-bearing marine deposits occur as high as 300 feet above present sealevel.[64] In this region there is no evidence of any postglacial sealevel higher than the present one. Probably the marine deposits are interglacial. The possibility has not been excluded that they owe their present altitude in part to crustal movement.

Pacific islands. Many islands in both the northern and southern regions of the Pacific Ocean bear emerged strandlines cut into bedrock. Persistent among these are strands said to record former sealevels at 5 feet and 25 feet above the present one, and, in addition, possibly other, still higher strands.[65] The strandlines are persistent throughout so wide a region that they have been regarded as eustatic. Possibly the 5-foot strandline is referable to the Climatic Optimum discussed in Chapter 21.

Europe and the Mediterranean. Pleistocene marine deposits and marine strandlines occur at many places along the Atlantic coast of Europe south of the glaciated region, as well as on both sides of the Mediterranean, ranging up to more than 300 feet above present sealevel. These features, which have been recorded by various observers in various districts, have never been continuously traced in the field. Attempts at correlating the published references have been made, notably by Depéret.[66]

Independent observations by R. V. Anderson[67] show four Pleistocene strandlines, some of them accompanied by well-developed wave-cut benches, along the Algerian coast. They are described as "fairly continuous" and maintain their horizontality throughout long distances. Altitudes of the inner and upper limit of each are 80 meters (264 ft.), 50 meters (165 ft.), 35 meters (115 ft.), and 14 meters (46 ft.).

Study of the Pleistocene deposits of the west coast of Italy, including both fossil-bearing marine deposits and terrestrial accumulations with

[64] P. S. Smith and Mertie 1930, pp. 238–241.

[65] Cf. Stearns 1941.

[66] Depéret 1926. A good summary in English of Depéret's correlations up to 1922, with critical comments by several American geologists, is given by Osborn and Reeds (1922). For a more recent critical summary see Blanc 1937.

[67] R. V. Anderson 1936, pp. 380–392.

plant remains, led to the recognition of three stages accumulated when the sealevel was higher than now, separated by stages of sealevel lower than now.[68] These stages, with their suggested relations to the glacial and interglacial stages, are shown in Table 26. The relations seem to be well established except those of the Calabrian, on which the evidence is obscure. The Calabrian stage has been held by Zeuner to include the First Glacial and even a part of the Pliocene. This correlation does not seem probable, but the matter awaits further study.

TABLE 26. PLEISTOCENE MARINE STAGES IN THE MEDITERRANEAN REGION

Sealevel (Position relative to present sealevel)	*Name of Stage*	*Fauna and/or Flora*	*Suggested Correlation with Glacial Stages*
Rising to present position	Flandrian	Fluctuating	Post-Fourth Glacial maximum
Low, with fluctuations		Cool, with fluctuations	Fourth Glacial
High	Tyrrhenian	Warm	Third Interglacial
Low		Cool-temperate?	Third Glacial
High	Sicilian	Warm, with cold zones	Second Interglacial
Low			Second Glacial
High	Calabrian	Warm, with cold zones	First Interglacial (?)

ASIA. The coastal plain of southwestern Kamchatka faces the Sea of Okhotsk on a 250-mile front between latitudes 52° and 55°. It is 10 to 30 miles wide, and its inner margin lies about 100 feet above present sealevel. This plain is described as consisting of two marine "terraces," of which the lower is 2 to 6 miles wide with an inner edge at less than 30 feet, while the higher reaches 100 feet.

AUSTRALIA. A strandline at about 100 feet has been reported from many segments of the coasts of Australia and Tasmania.[69]

Evidence of Lower Sealevels

Evidence of former sealevels lower than the present one is of several different kinds. The various lines of evidence, and the inferences that have been drawn from them, are indicated below.

SUBMERGED STRANDLINES. Strandlines below present sealevel are very

[68] Blanc 1937; Zeuner 1938. (Table 26 is based chiefly on Zeuner. Unpublished information indicates that Zeuner has since introduced changes, received so recently that they can not be represented here.)

[69] Browne 1945, p. xii.

difficult to identify, for at least two reasons: First, the times during which the sealevel paused at one position during a glacial age were, because of fluctuation in the volume of the glaciers, individually short. Further, strandlines fashioned in the weak sedimentary materials that cover the continental shelves are rather easily destroyed by the sea after submergence. It is not surprising, therefore, that only one such feature has been identified with much confidence. This is the *Franklin shore,*[70] a gentle-scarplike break in the seaward slope of the continental shelf off eastern North America. It extends almost continuously from the latitude of Philadelphia to the latitude of lower Chesapeake Bay, a distance of nearly 200 miles, and may extend farther south through a sea-floor area as yet uncharted. Its southern end, as far as it has been charted, lies at a depth of 240 feet; its northern end at 330 feet. Thus it appears to have been tilted down toward the north. But as its southern end lies north of the northern end of the Surry scarp we can not infer that because it is tilted it antedates the Surry scarp. Although Veatch and Smith attributed the tilt to incomplete recuperation from glacial loading, it seems unlikely that this loading affected the crust as far beyond the drift border (nearly 300 miles) as such an interpretation would imply.

A somewhat similar break in the submarine slope off the massive delta of the River Rhône, at a depth of slightly less than 100 meters, has been recognized[71] with the suggestion that this may be a strandline made during the Wisconsin Glacial age. The possibility of crustal warping makes it unsafe to infer that the present depth of this break results solely from rise of sealevel.

At Nome, Alaska, surface features and buried beach placers record several marine strandlines, apparently horizontal, lying at various positions from 34 feet below to 79 feet above sealevel.[72] At least two marine transgressions seem to be indicated, apparently related to eustatic changes.

Submerged organic remains. At several places around the Bay of Fundy there are extensive occurrences of tree stumps rooted in positions at least 15 feet and perhaps as much as 59 feet below present mean sealevel.[73] The trees, some of which have 200 annual growth rings, are well preserved and include spruce, hemlock, pine, beech, birch, alder, poplar, elm, ash, and tamarack. In themselves these stumps prove only that the land in these places has been submerged, but, because there is clear evidence that Nova Scotia was differentially warped *up* during the last

70 Veatch and Smith 1939, p. 44.
71 R. J. Russell 1942, p. 232.
72 Moffit 1913, pp. 40–49, 109–124.
73 J. W. Goldthwait 1924, pp. 155–161.

deglaciation, it seems almost certain that these forests were submerged specifically by rise of sealevel against the already warped-up land.

Less conspicuous but none the less definite occurrences of tree stumps *in situ* and of terrestrial peat are found elsewhere along the Atlantic coast in New England and as far south as Carolina. Along the New England coast the organic remains clearly postdate the latest glaciation of the district. At Boston a submergence of at least 43 feet is recorded.[74] Submerged peat containing birch and willow is exposed on the coast of Iceland, indicating post-peat submergence of at least 13 feet.

Similarly, submerged peat beds, well-preserved stumps *in situ,* and prostrate tree trunks occur in places off the whole length of the shoreline of England and Wales, and on the Dogger Bank in the North Sea at depths locally as great as 170 feet.[75] The trees include birch, willow, hazel, and oak. From the shallow floor of the North Sea there have been dredged up the bones or teeth of woolly mammoth, musk-ox, reindeer, woolly rhinoceros, and hyena.

The Long Forties Bank off the east coast of Scotland lies in a deeper part of the North Sea, but its top, covered by only 200 to 250 feet of water, is floored with pebbles and broken shells, including mollusk species that live only in the zone between high and low tide.[76]

Although the relative dates of the bones, shells, peat, and trees are not known, the animal remains combine with the plants to show that fairly recently a submergence of at least 170 feet has taken place, and that this submergence is probably the result of rise of sealevel since the maximum of the Fourth Glacial age. In Europe the name *Flandrian* has been given to the marine deposits laid down since the maximum of the latest marine regression, coinciding with the last glacial maximum.[77]

Making use of earlier-published data on the depth of habitat of certain mollusks occurring as fossils in the Flandrian clays in southern Norway and on the presence of terrestrial plants on the North Sea floor, Farrington[78] concluded that the rise of sealevel would be far greater than that arrived at in most of the calculations on record. The method is interesting, but it may be questioned whether the depth at which a certain mollusk now lives in the Arctic Sea is the same as the depth at which the same species lived in a certain part of the North Sea at a former time. Modern ecologic investigations suggest that depth of habitat is a product of temperature, turbidity, sunlight, aeration, and similar underlying controls. From this follows the possibility that a species living at a given

[74] Shimer 1918, p. 457.
[75] Godwin 1943, Clement Reid 1913.
[76] Jukes-Browne 1892, p. 402.
[77] Georges Dubois 1930.
[78] Farrington 1945.

depth at a given time and place may live at a different depth at another time and place. Therefore caution seems necessary in basing estimates of deleveling on this factor.

Submerged deltas. Lower sealevels should be indicated by deltas, especially those built of glacial outwash entering deep water, and many of these should be recognizable by inspection of detailed charts. Submerged deltas have thus been tentatively identified on the south coast of Newfoundland.[79] A massive delta fan, now submerged, occurs at the mouth of the River Rhône.[80] The general occurrence of large submerged deltas has been discussed by Shepard.[81] The fact that a delta is submerged does not prove rise of sealevel, because a bulky delta may have enough weight to cause crustal subsidence or may be built in a subsiding area.

Distribution of sediments on shallow sea floors. In normal circumstances the sediments on shallow sea floors decrease in grain size with increasing distance from the shore — the source of supply — and with increasing depth of water. On the shelf off eastern United States, however, this ideal offshore gradation from coarse to fine sediment is only partly realized. On some parts of the shelf, sediments far from shore are notably coarser than those inshore. Furthermore they include windblown sand, recognizable by its rounded and frosted grains, at depths as great as 180 feet.[82] These facts are at least in considerable part attributable to the lower sealevels of the glacial ages.

Drowned stream valleys. The shallow sea floors in various parts of the world are engraved with channels whose forms have been commonly attributed to rivers flowing on the land. Accordingly it is believed (although we can hardly say it is proved) that these channels were cut during the lowered sealevels of the glacial ages and that their submergence is so recent that marine sediments have not yet succeeded in filling them up. One of the earliest-described and best-known of these channels is the one that continues the course of the Hudson River southeastward across the continental shelf. The channel is 90 miles long and ends at a depth of about 240 feet.[83] Some of the rivers of western Europe have submerged channel prolongations. Notable among them are the former extensions of German and English rivers across the floor of the North Sea. The Elbe channel can be traced as far north as the latitude of Edinburgh, and the Rhine channel, evidently joined by the major streams of eastern England, is visible even farther north and extends

79 Flint 1940*a*, p. 1773.
80 Baulig 1927.
81 Shepard 1928.
82 Stetson 1938.
83 Veatch and Smith 1939, chart 5.

down to a depth of at least 300 feet. These features are clearly evident on a contoured map of the North Sea floor.[84]

The Bosporus and Dardanelles straits are the submerged channel of a stream that during one of the glacial ages drained a vast lake of meltwater in the Black Sea basin into the glacially lowered Mediterranean Sea.[85] Still another instance of a drowned river valley system is found on the Sunda Shelf, between Borneo and the Malay Peninsula, extending down to a depth of 300 feet.[86]

The amazing "seavalleys" that furrow the great continental slopes and extend down to depths of nearly 2 miles[87] should not be confused with the channels at shallow depths on the continental shelves. Some writers, including F. P. Shepard and C. Johns, have attributed such features to river cutting during a vast lowering of sealevel caused by the building up of enormous hypothetical ice sheets. But the evidence against this view is so weighty that the idea has been dismissed by nearly all geologists. Probably most of the great seavalleys are the work of erosion that occurred beneath the level of the sea.

The floors of large valleys in many parts of the world lie so far below sealevel that it is probable they were excavated at times when the sealevel stood lower relative to the land than it does now. In Britain a number of major drift-filled valleys have bedrock floors 150 to 200 feet below sealevel.[88] According to David the valleys of major Australian rivers have been drowned in recent time to depths of 150 to 200 feet. "This may have been due in part," he wrote, "to a rise in the ocean level following on the melting of the Pleistocene ice sheets."[89] In southern Africa borings made near the mouths of large valleys show that these valleys have been submerged and filled with sediments to depths exceeding, in one valley, 150 feet below present sealevel.[90]

Borings made on various parts of the coast of northern New England[91] and along major valleys in Connecticut show that the bedrock floors of the river valleys lie much below present sealevel.

It can not be said that these depths are wholly the result of rise of sealevel. The whole of New England was glaciated, and glacial erosion probably played a part in deepening the valleys. The reverse gradient of the bedrock floor of the Connecticut between East Hartford and

[84] See R. G. Lewis 1935, map.
[85] Wilhelm Ramsay 1930, p. 40.
[86] Molengraaff and Weber 1921.
[87] Cf. Veatch and Smith 1939.
[88] Slater 1929*a*, p. 489.
[89] David 1932, p. 96. See also summary of other data in Browne 1945, p. xiii.
[90] Du Toit 1939, p. 406.
[91] F. G. Clapp 1907, p. 513.

VALLEY	LOCALITY	MILES ABOVE MOUTH	ALTITUDE OF ROCK FLOOR OF VALLEY (feet below sealevel)
Housatonic	Stratford, Conn.	4	90
Quinnipiac	New Haven, Conn.	1	140
•	Plainville, Conn.	28	28
Connecticut	Lyme, Conn.	1	180
	Middletown, Conn.	26	118
	East Hartford, Conn.	40	More than 209
	Thompsonville, Conn.	56	More than 116
Thames	New London, Conn.	3	More than 215
Charles	Boston, Mass	Various	200–400
Merrimac	Amesbury, Mass.	6	100
Fore	Portland, Me.	1	110
Androscoggin	Brunswick, Me.	18	155

Middletown strongly suggests this factor. The bedrock floor beneath the Hudson River shows an even more pronounced reverse gradient. In both the Connecticut and the Hudson, however, the geologic conditions are exceptionally favorable for effective glacial erosion. As these conditions are not shared by the other places shown in the table, and as similarly drowned and partly buried valleys are common in regions that were not glaciated,[92] probably the observed submergence and burial are chiefly the result of rise of sealevel.

WINDBLOWN SANDS WITH INADEQUATE SOURCE OF SUPPLY. The island of Bermuda is built of sedimentary rocks resting (well below present sealevel) on a broad volcanic platform. The sedimentary rocks consist chiefly of lithified windblown sands (*eolianites*), the sand grains being tiny particles of limestone and marine shells. The present island is surrounded by a broad shelf or bank far larger than the island itself and covered in most places by less than 30 feet of water. Under existing conditions windblown sands can not accumulate because there is no source of sediment. But, with sealevel lowered by only a small amount, a vast supply would be made available by the emergence and drying out of the bank. The inference is plain that the limy sediments were heaped into dunes during the lowered sealevels of the glacial ages, for there is evidence that the Bermuda mass as a whole has not moved significantly during Pleistocene time. Furthermore there have been recognized five distinct eolianites separated from each other by zones of weathering.[93] Evidently these record glacial ages or sub-ages or both, but their exact dating is uncertain.

Somewhat similar dune sands having no source of supply under

[92] For example, the floor of the Delaware River valley beneath Pleistocene sediments extends down to more than —50 ft. at Bordentown, —50 ft. at Camden, —84 ft. at Wilmington, and —110 ft. near Cape May. (Figures kindly supplied by Meredith E. Johnson, State Geologist, New Jersey.)

[93] Sayles 1931.

present conditions and probably referable to glacially lowered sealevels occur in the Bordeaux district of France, and in Iceland.[94]

Anomalous faunal distributions. The similarity of faunas, especially mammal faunas, on pairs of land areas separated by shallow water implies the recent presence of connecting land bridges and thus provides evidence in support of the concept of lowered sealevels within a comparatively late part of Pleistocene time. The cosmopolitan early-Pleistocene mammal fauna common both to the mainland coast and to some islands of the Mediterranean are an example.[95] Again, Britain has only about half as many species of mammals, reptiles, and amphibians as has the adjacent continent, while Ireland has only about half those present in Britain.[96] As these animals must have come from the mainland since the maximum glaciation, two land bridges are implied: one at the English Channel and one between Britain and Ireland. A reduction of sealevel of 150 feet would create both bridges. The disparities in the number of species suggest that the bridges did not last long after conditions favorable for migration came into existence.

It is well recognized that some of the Pleistocene mammals of North America came from Asia and that a land bridge at the Bering Strait was their most probable path of entry. Emergence of only 180 feet would now create such a bridge across this 50-mile intercontinental gap between peninsulas of low land. Although some large animals might have crossed the strait on sea ice such as forms there during the winters today, it is thought unlikely that many small animals did so. The region of the Strait exhibits no evidence of having been glaciated at any time during the Pleistocene or of having experienced crustal movement since the time of the last glaciation. Thus the evidence, as far as it goes, suggests that, when the sealevel was lowered during the glacial ages, this process alone was adequate to create a land bridge between Asia and North America at the Bering Strait.

The close similarity in the faunas and floras of Victoria (Australia) and Tasmania is attributed to glacial-age emergence of the shallow sea floor that separates them. An emergence of less than 250 feet would connect Tasmania to the Australian mainland. An analogous relationship exists between the islands of Borneo and Sumatra, whose faunas are closely similar. This evidence confirms the belief, based on the drowned valleys in the intervening Sunda Shelf, that these islands were recently connected. Probably the connection dates from the lowered sealevels of the glacial ages.

There are other examples of faunal similarities between lands sepa-

[94] Daly 1934, pp. 197–199.
[95] Blanc 1937.
[96] Jukes-Browne 1892, p. 402.

rated by shallow seas. But before any one of them can fairly be ascribed to lowered sealevels the unlikelihood of crustal movements as the cause of present submergence must be shown. It is difficult if not impossible to exclude crustal movement entirely. In consequence a judgment of "probable but not proved" must be entered in the case for temporarily lowered sealevel.

In summary, we can say that proof exists that within the limits of Pleistocene time the sea has stood both higher and lower *with respect to the land* than it does today. Further, we can point to strong evidence that these fluctuations are the result of changes in the level of the sea itself, brought about by the building and wasting away of glaciers on the land. But we have to admit that crustal movements have played an important part in Pleistocene changes of level, and that as yet we lack data adequate to enable us to separate clearly in each place the effects of shift of sealevel from the effects of crustal movement.

Effect on Stream Regimens

Unquestionably the rising sealevels brought about by change from glacial to interglacial conditions drowned the mouths of large streams in many parts of the world. Unquestionably also the falling sealevels at times when glaciers approached their maxima laid bare vast areas of the continental shelves, which were trenched and channeled by the extended rivers of those times. Along considerable stretches of some rivers the general effect of these slow fluctuations would have been twofold: (1) filling of valleys with sediment by slackened streams, and (2) erosion of valleys by accelerated streams. But this statement does not imply that fluctuations of sealevel alternately slacken and accelerate all streams. The response of a stream to a change of sealevel depends on the shape of its long profile, the volume and grain sizes of the load of sediment it is carrying, the configuration of the sea floor off the stream's mouth, the local stability of the Earth's crust, and other factors. It follows that no two streams will respond in just the same way to fluctuations of sealevel. Accordingly it is unsafe to infer changes of sealevel solely from evidence of cutting and filling by streams, as has unfortunately been done by writers in both Europe and North America. Evidence of changes of stream regimen may properly be shown to be in agreement with the scheme of sealevel fluctuations inferred from marine deposits, strandlines, and similar direct evidence, but it does not, at least in the present state of our knowledge, in itself constitute proof of movements of sealevel.

Chapter 20

GLACIAL AND INTERGLACIAL CLIMATES

INTRODUCTION

The fluctuation of climates within the Pleistocene epoch has been far more complex than was realized a generation ago. The stratigraphic record of alternating glacial and interglacial ages within the glaciated regions and the record of changes in the levels of lakes within the nonglaciated regions clearly imply long-enduring climatic fluctuations of considerable amplitude. But the evidence is fragmentary and of several quite distinct kinds. Accordingly we shall try to set it forth in sections, keeping in mind that when we have finished we shall have a picture so incomplete that it ought perhaps not to be called a "picture" at all.

EXTENT OF GLACIERS

By many the "Ice Age" is loosely thought of as having been characterized by a vast glaciation, whereas the present time, in contrast, is thought of as one in which glaciers are insignificant. Yet against this view are the following facts:

1. During the maximum extent of glaciers (regardless of date) about 32 per cent of the present land area of the world was covered by glacier ice.
2. During the latest glacial age about 27 per cent of the land was ice covered.
3. At present about 10 per cent of the land area is ice covered.
4. During the interglacial ages (with combined length greater than that of the glacial ages), and in preglacial time, little or no glacier ice existed on the lands (except, perhaps, on the Antarctic Continent, about which so little is known that any figure would be misleading).

These facts lead to the belief that the present day, though transitional, is glacial rather than interglacial and probably should be considered a part of the Wisconsin Glacial age. Therefore the present can not be considered a standard against which to measure the data of a glacial climate; the present itself has so many glacial characteristics that such a comparison would not be justified. However, the fact that in one glacial

age there was nearly three times the area of glacier ice that now exists strongly suggests that during that time climatic contrasts were considerably greater than they are at present.

THE GLACIAL AGES CONTEMPORANEOUS THROUGHOUT THE WORLD

Throughout the world we have not found a single place where the glaciers of today are more extensive than the glaciers of the past. Everywhere glaciers have been more extensive formerly than they are now. However, this fact does not prove that the glacier expansions took place contemporaneously throughout the world. Other evidence must be sought.

There is little doubt that the successive glaciations were synchronous throughout the northern hemisphere. The continuity of the Laurentide drift and its interfingering with Cordilleran drift establish the fact within North America. In Eurasia similar continuity exists from the Netherlands and Denmark eastward to the Yenesey River in Siberia. Farther east in Siberia the latest glaciation is said to have been synchronous from the Lena River to Bering Strait, although the region has admittedly been examined only in reconnaissance.

In comparing the northern hemisphere with the southern, we can not say so definitely that the glaciations of the two were synchronous. Geologists familiar with the drifts of both polar hemispheres agree in the opinion that the latest glaciation was contemporaneous in them. This opinion is based on the degree of alteration of the drifts, on the freshness of the drift topography, and on the apparently consistent relation between the regional snowline of the present and that of the glacial maxima. Admittedly this does not constitute proof. It is subjective, but it is a considered opinion. And at any rate it is all we have at present. It is strengthened somewhat by the clear evidence of rise and fall of sealevel through hundreds of feet during the Pleistocene epoch. If glaciation had alternated between the northern and southern hemispheres the fluctuation in sealevel would have been comparatively small. It is strengthened further by the record of large fluctuation of the East African lakes discussed later in this chapter. Lying close to the equator these lakes should have experienced only small fluctuations, or none at all, if the glaciations had alternated.

In addition we may cite the fact that corresponding climatic fluctuations postdating the maximum of the Fourth Glacial age seem to be recorded in North America, Eurasia, East Africa, South America, Australia, New Zealand, and the Antarctic Continent. This fact includes

correspondence, in all these regions, in the regimens of the existing glaciers, which have been shrinking notably during the last few decades,[1] as well as worldwide rise in temperature during the same period, as outlined in Chapter 21.

Nordenskjöld inferred contemporaneous glaciation of the two polar hemispheres from faunal evidence, which, although suggestive, is far from conclusive.[2]

Finally nowhere in the world is there reported an existing glacier, even at the equator itself, that does not lie within an area of formerly much more extensive glaciation of comparatively recent date.

As between the north and south polar hemispheres, therefore, a question still exists, though the evidence points towards synchrony. As between the eastern and western hemispheres synchrony can be regarded as established.[3] As future research succeeds in reaching and studying the sediments beneath the sea floor it may become possible to trace glacial and interglacial marine deposits directly between continents and between hemispheres. Another promising line of inquiry lies in the former positions and relative dating of the regional snowline in equatorial mountains. Admittedly the evidence will be fragmentary. But from the altitudes of cirques and of the lower limits of glaciation in these mountains it should be possible to determine whether the regional snowline rose from one polar hemisphere into the other or whether it rose toward the equator from both hemispheres at the same time.

No one of these arguments is conclusive, but the weight of evidence is strongly suggestive that the glacial ages were synchronous throughout the world.

DEPRESSION OF THE REGIONAL SNOWLINE

There is no doubt whatever that at the maximum of the latest glacial age the regional snowline stood lower than it does now. This has been found true wherever measurements have been made.[4] As is stated in Chapter 23, reduction of the snowline was primarily secular but was somewhat exaggerated on each mountain range by the local presence of new or increased snow and ice.

[1] Cf. Thorarinsson 1940.

[2] Nordenskjold and Mecking 1928, p. 41.

[3] References on this subject are few and rather unsatisfactory. See Antevs 1928*a*, p. 13; Daly 1934, pp. 30–41; Woldstedt 1929, p. 283.

[4] It was stated in Chapter 6 that the average altitude of the floors of abandoned cirques in any district give a rough approximation of the minimum altitude of the regional snowline at the time (or rather times) when the cirques were made. (Cf. Penck and Bruckner 1909, *I*, p. 266). A related line of evidence is the altitude of the lower limit of glaciation at the crest of a mountain range only part of which has been glaciated (Sharp 1938, p. 311). The glacial-age snowline in western United States is described in Chapter 12.

In contrast, one of the valleys of the Alps, near Innsbruck, has yielded evidence of a regional snowline higher than now. The Hötting deposits, belonging to one of the interglacial ages, include six species of plants that no longer grow in the Alps but whose present habitat indicates a snowline some 1300 feet higher than the present snowline in the district where the fossil plants are found.[5]

Thus we have evidence, fragmentary and not very exact to be sure, that during the Pleistocene epoch the regional snowline has stood both lower and higher than it does today. In the Alps the figures given are about 1300 feet higher and about 4000 feet lower than now, a total range of fluctuation of about 5300 feet.[6]

The amount of depression of the regional snowline at the maximum of the latest glacial age varied from place to place, and the variations were consistently related both to distribution of precipitation and to latitude. Wherever measurements have been made in mountain regions, it has been found that, for a given latitude, the greater the precipitation the greater was the amount of depression of the regional snowline during the last great glacial maximum. The mountains of western United States illustrate the point. In the Glacier Park district of Montana the depression of snowline was considerably less than on Mount Rainier, one of the high peaks of the Cascade Mountains in Washington. The two places are at nearly the same latitude, but the precipitation on Mount Rainier is two to four times as great as in the Glacier Park district. A similar relationship exists between the Rocky Mountains in Colorado and the Sierra Nevada in the same latitude belt. In fact the principle involved applies wherever snowline measurements have been made.

On the other hand, the depression of snowline on Mauna Kea on the Island of Hawaii was less than half as great[7] as that on Mount Rainier, although the altitudes of the two peaks and the annual precipitation figures for their upper slopes are rather closely comparable. But there is a difference of 27 degrees of latitude between them. Thus latitude as well as snowfall controls the difference between the present snowline and the depressed snowline of the last great glacial maximum.

Neglecting conspicuous differences in precipitation, we can say that the snowline depression was greatest in the middle latitudes (say between 45° and 65°) and that from those latitudes it diminished both toward the poles and toward the equator. The small depression in the equatorial region is explained to a large extent by strong summer ablation. The

[5] Penck 1921.
[6] Summarized in W. B. Wright 1937, p. 166.
[7] Perhaps 2000 feet (Wentworth and Powers 1941, p. 1201).

small depression in the polar regions is explained partly by relatively small precipitation and partly by the fact that even today the snowline is already so low that a very slight depression would push it down to sealevel.

The measured depression of snowline (below its present position), affected both by latitude and by precipitation, amounted roughly to 1200–2000 feet in the polar regions and in the mountains of low-latitude deserts, 2000–3000 feet on equatorial mountains, and 3000–4000 feet in the middle latitudes.

The remarkably great depression of snowline in the middle latitudes was brought about in part by an additional factor. In the Alps the regional snowline lies at a higher altitude than the zone of maximum snowfall. But in the mountains of northern Scandinavia these relations are reversed, because the regional snowline descends toward the north more steeply than does the zone of maximum snowfall.[8] Where the two lines intersect, the northward inclination of the regional snowline steepens, because summer ablation is sharply reduced north of the intersection.

The zone of intersection of these two lines falls between 55° and 65° N latitude. This is the zone in which the great Scandinavian ice sheets of the glacial ages originated. It is also the zone of origin of the vast Laurentide Ice Sheet of the Wisconsin Glacial age, and probably of its predecessors in the earlier glacial ages.

It is not yet possible to determine whether conditions in the North American Cordillera between the United States and Alaska were similar, because adequate snowline data are not available, but it is probable that an analogous relationship exists.

Calculated Reduction of Mean Temperature

Penck and Brückner calculated that at the maximum of the Fourth Glacial age the depression of snowline on a number of ranges in the Alps averaged 3300 feet below that of today, and that depression of this amount could have been brought about if the mean annual temperature had been about 5° C less than it is now. Using a similar method Penck[9] calculated that in coastal and western Europe and in Italy and Greece the mean annual temperatures at the same time were 7° to 8° C lower than they are at present. These figures of course can not be exact, but probably they approximate the truth. The reduction of temperature

[8] Penck and Brückner 1909, p. 1144; Paschinger 1923.
[9] Penck 1936*a*, p. 226.

throughout the world during the glacial ages was certainly not uniform. It varied from place to place with many local influences.

It is interesting to compare the differences between glacial-age temperatures and present temperatures as calculated by various authorities from various lines of evidence. Klute[10] and Antevs,[11] using Klute's data, have made estimates based on snowline figures for mountains in western North America, amounting to 4° to 5° C and 4.4° C respectively. Meyer arrived at a corresponding figure of 3° to 4° C for the equatorial Andes.[12] Hume and Craig derived values of 5° to 7° C for the high mountains of East Africa.[13] Meinzer, basing calculations on the areas of present and glacial-age lakes in the Basin-and-Range region of western North America, made a rough estimate of 8° C.[14] Endo calculated from plant evidence that glacial-age temperatures in central Japan were lower than now by 5° to 6° C.[15]

It is noteworthy that these figures, when compared, are not inconsistent. Those in mountain regions are lower than those for coastal and plains regions; such a result is expectable.

Penck and Brückner argued that cirques and other snow-catchment basins in the high ranges of the Alps could have held no more snow during the glacial maxima than they do today, and they inferred that greater former glaciation of the Alps was the result of decreased temperature entirely, rather than of increased precipitation.[16] However, as pointed out by Huntington, the evidence cited by Penck may be taken equally well to indicate that the zone of maximum snowfall was shifted downward as a result of lowered mean temperature, so that the highest mountains stood above this zone. Then the precipitation during the glacial maxima could have been, and probably was, greater than now, though it would have been concentrated at altitudes below the highest parts of the Alps.[17]

CHILLING OF SEA WATER; SEA ICE

Climatic fluctuation during the Pleistocene is reflected in fluctuation of temperatures in the sea, which in turn is revealed by alternations of "cold" and "warm" fossil marine faunas and by alternations of organic

[10] Klute 1928, p. 70.
[11] Antevs 1935, p. 307.
[12] Meyer 1904.
[13] Hume and Craig 1911.
[14] Meinzer 1922.
[15] Endo 1933, p. 179.
[16] Penck and Bruckner 1909.
[17] Huntington 1925, p. 185.

deposits and layers of sediment believed to have been deposited from abundant floating ice.

Direct evidence of chilling of Pacific waters off southern California, a thousand miles from any tidewater glaciers, is afforded by an extraordinary "cold zone" in the Santa Barbara formation, containing fossil invertebrates that appear to record a mean water temperature about 10° C lower than now.[18] Conversely, the interglacial Gardiners clay of Long Island carries a fauna that indicates water slightly warmer than now. The heightened temperature of the coastal waters of southern New England reflects the disappearance of glacier ice and, with it, diminished vigor of the Labrador current which today brings cold water to that region.

The floor of the Arabian Sea (Indian Ocean) at about 10° N latitude has yielded sediments in layers consisting of pelagic Foraminifera of alternating warm- and cold-water types.[19] This fact strongly suggests climatic fluctuation, but the evidence does not yet permit correlation with specific glacial and interglacial time units.

Similar stratification has been observed in sediments from the floor of the South Atlantic, with a superficial zone containing warm-water Foraminifera overlying a zone containing cool-temperate types.[20] The lengths of time represented by these layers, however, have not been estimated.

A suite of 35 core samples from the continental slope off the east coast of the United States between Cape Cod and Cape May revealed a double alternation between a pelagic fauna of warm, Gulf Stream type and a fauna of Arctic and subarctic type.[21]

The most remarkable stratigraphic section obtained from the sea floor is revealed by eleven cores, up to 10 feet in individual length, taken in a traverse between Newfoundland and Ireland near 50° N latitude.[22] The cores consist of alternating layers of glacial-marine deposits (gritty and stony material dropped from floating ice, with diminished calcium carbonate and a sparse fauna), alternating with foraminiferal marl like that forming today. This alternation is believed to represent an alternation of glacial and nonglacial climates, but whether the individual layers

[18] Crickmay 1929, p. 634. This zone is tentatively correlated by Barbat and Galloway (1934) with the basal San Joaquin clay in interior California, which they regard as upper Pliocene. Pleistocene and upper Pliocene correlation in California has not yet been widely agreed on.

[19] Stubbings 1939. The longest core obtained shows four "cold zones." An attempt to relate the stratigraphy to the Swedish chronology obtained from varved sediments, discussed in Chapter 18, is interesting but unconvincing.

[20] Schott 1935.

[21] Phleger 1939.

[22] Bradley and others 1940.

have the value of stages, substages, or some other subdivision is not yet known. Four glacial-marine and four foraminiferal layers appear in the cores, but it is not known how many more may underlie the known sequence. Correlation of the eleven cores with each other is facilitated by the presence of a persistent, recognizable zone of volcanic ash.

Three core samples of ooze from the floor of the Caribbean Sea south of Cuba, though naturally including no glacial-marine sediment, nevertheless reveal alternations of Foraminifera that suggest fluctuating changes in water temperature.[23]

Sea ice forms today when air temperatures are reduced sufficiently to freeze the surface water of the sea. At present freezing occurs in winter throughout most of the Arctic Sea, through a broad belt surrounding the Antarctic Continent, and in parts of smaller water bodies such as Bering Sea and the Sea of Okhotsk. Sea ice is normally 7 to 12 feet thick. It is cracked and broken up by marine currents and gradually drifts into the warmer waters of lower latitudes, where it melts.

Whether sea ice existed in extreme high latitudes in interglacial and preglacial times is not known. During such times it probably was at least much less extensive than now, especially in the Arctic Sea. This seems likely because very slight climatic fluctuations in modern times appear to result in very great changes in the amount of sea ice produced; hence the much greater changes from glacial to interglacial conditions should have had correspondingly greater effects.

Once sea ice is formed over a large area, it reflects far more solar heat than did the water which it replaces, causes chilling of the atmosphere above it, and tends to create anticyclonic conditions such as characterize the ice-filled Arctic Sea today, thus affecting regions outside the actual area of sea ice. Sea ice is important also in reducing the amount of moisture that can be absorbed by air masses passing over the water body. Thus the moisture content of polar air masses should have been more abundant for nourishing the glaciers of Arctic North America and Eurasia during the earliest phase of a glacial age, when presumably the Arctic Sea was still unfilled or only partly filled with sea ice, than during the height of a glacial age when it must have been completely filled with floating ice.

It seems likely that at the heights of the glacial ages sea ice filled perennially not only the Arctic Sea but also the Norwegian or Greenland Sea in the Atlantic sector and the Bering Sea in the Pacific sector. It is not likely that perennial ice extended far south of the 60th parallel in mid-Atlantic or far south of the Aleutian Chain in the Pacific region because warm currents would soon have melted it (Plate 3).

[23] Cushman 1941; see also Pigott and Urry 1942, p. 1199.

FREEZING AND THAWING OF THE MANTLE

Present-Day Frozen Ground

Winter freezing of the soil is a matter of common observation to farmers throughout wide regions in middle latitudes. But in the northernmost lands of both North America and Eurasia the mantle, where it is unusually thick, is perennially frozen down to depths that amount in places to hundreds of feet. Freezing has occurred because under the existing climatic conditions the warmth of the short summers has been inadequate to supply as much heat to the ground as is abstracted from it during the long cold winters. In summer, shallow surface thawing occurs to depths of a few inches to a few feet, but below this very superficial zone the ground remains frozen throughout the warm season. In general the limit of continuous frozen ground lies close to the mean annual isotherm of 0° C.

The frozen part of the mantle is known in Sweden as *tjaele,* a term that appears to be coming into international use. In summer the tjaele stops the normal downward percolation of subsurface water derived from thawing of the soil above it and from precipitation. As a result the thawed mantle above the tjaele becomes saturated with water and flows downslope with the tjaele serving as a hard floor beneath it. Flow of this kind, usually termed *solifluction,* occurs on even the gentlest slopes. During the process the flow earth becomes thoroughly mixed, and as a result it is not stratified. If the mantle was stratified originally, stratification in the superficial zone is destroyed.

Where the ground is flat and nearly horizontal, seasonal freeze and thaw sort the materials above the tjaele into stripes and polygonal networks of stones separated by finer materials, chiefly silt and clay. The polygons, usually many feet in diameter, give a strikingly artificial appearance to the landscape. The mechanism by which this sorting occurs[24] is not thoroughly understood and is not necessary to the present discussion, which aims only to state what is now commonly agreed, namely, that perennially frozen ground and the accompanying processes of solifluction and the sorting of stones into polygons and stripes are associated with climates in which the mean annual temperature is less than 0° C.

Evidence of Former Frozen Ground

Former perennially frozen ground is recognized by the peculiar stone stripes, polygons, and flow earth that develop above the tjaele during

[24] See discussion by Sharp (1942*b*) and references listed therein.

the summer thaw. Because these features occur today in an "inactive" condition, that is, not being formed or added to, in regions too warm for significant freezing of the ground, it is concluded that the climates of such regions were formerly colder than now. It is highly probable, furthermore, that the former colder conditions occurred during glacial ages. Conversely, as is shown below, clear evidence of former thaw is found in at least one region in which the ground today is firmly frozen.

Thus, although stone polygons appear to be forming today in the Alpine zones of high mountains such as the Presidential Range (at altitudes over 5000 ft.) and Mt. Monadnock in New Hampshire (3166 ft.), inert or "fossil" polygons are reported from Mt. Katahdin (5267 ft.) and Mt. Desert Island (1100 ft.) in Maine, as well as in the Scottish Highlands and the Lake District in northeastern England.[25]

The flow earth which is the peculiar product of solifluction has been recognized in Britain, France, northern Germany,[26] and Denmark.[27] In Britain flow earth is widely known as *head*.[28] Distinct sheets of head are identified with drift sheets of at least two different ages; the older heads are weathered and much dissected. Head does not seem to be forming today except near the tops of conspicuous highlands.[29]

Sheets of flow earth occur extensively in the Somme Valley in northern France, where they have been studied in detail.[30] Many artifacts have been recovered from the Somme Valley flow-earth layers, which it is claimed can be correlated by means of these finds with four glacial stages recognized elsewhere in Europe. Former mild frost heaving has been recognized as far south as Bordeaux,[31] where the adjacent cold, ice-studded Bay of Biscay may have helped to keep the temperature low. However, the localities at which flow earth and other evidence of frozen ground have been reported are generally confined to a belt about 150 miles wide beyond the edge of the former Scandinavian Ice Sheet. No doubt the cold radial katabatic winds and the absorption of heat by melting at the margin of the ice sheet, as well as persistent anticyclonic weather conditions, were instrumental in bringing about low temperatures in the peripheral belt.

In north-central Illinois involutions in stratified fine sediments have been considered to be an incipient form of flow earth developed during

25 Denny 1940, p. 432.
26 Beschoren 1932, p. 79; pl. 4, 5.
27 Madsen and others 1928, p. 85.
28 A term first used by De la Beche in 1839.
29 Dines and others 1940.
30 Breuil 1934.
31 Bastin and Cailleux 1941.

late Cary or early Mankato time.[32] Similar involutions have been recognized in southern Connecticut.[33] "Fossil" stone streams occurring in the Driftless Area of southern Wisconsin have been regarded as having been formed during the Wisconsin Glacial age.[34]

Not infrequently involutions and flow earth are associated with windblown silt and with ventifacts, which are generally thought of as the result of the greatly increased wind strengths in the belt surrounding an ice sheet.[35]

Accelerated flowing and sliding of the mantle, not necessarily related to perennially frozen ground, have been reported from areas far from the limits of the former ice sheets. In the mountains of northern New Mexico two episodes of landsliding of early Pleistocene till have been referred to the cold climates of two later glacial ages.[36] In northwestern South Carolina a zone of peaty matter has been uncovered by the removal of a thickness of 20 feet of overlying soil material that has crept downslope as a result of mass-wasting. The peaty matter contains the pollen of both fir and spruce, which do not now grow in the vicinity. The overlying material appears to have crept downslope at a time when the climate was colder than now, perhaps during a glacial age.[37]

Evidence of Former Thaw

The examples cited indicate climates formerly colder than now. The reduced temperatures resulted partly from shifting of the subarctic climatic zone toward the equator and partly from the cold conditions induced more or less locally by the growth of large ice sheets. On the other hand there is on record one evidence, of the same general kind, of a former climate significantly warmer than the present one.

Central Alaska, a region that largely escaped glaciation because of its low altitudes, small precipitation, and warm summers, is widely underlain by thick deposits of silt whose origin appears to be alluvial, colluvial, and eolian. Most of this silt lies within the Arctic belt of perennial deeply

[32] Sharp 1942*a*.

[33] Denny 1936.

[34] H. T. U. Smith 1941.

[35] The association of flow earth, loess, and ventifacts has been considered as indicating "periglacial" conditions, by which are implied very cold, windy conditions in the neighborhood of an ice-sheet margin. This general term can have no absolute quantitative significance; strictly speaking, a periglacial climate is any climate that exists around the edge of a large glacier and might be of more than one kind. Progressive change of periglacial climate has been specifically recognized by Cailleux (1938).

[36] H. T. U. Smith 1936.

[37] Eargle 1940.

frozen ground. Sections of this silt exposed by mining operations record an interval of deep thaw accompanied by extensive erosion. Obviously this interval, which occurred between times of deep and long-continued freezing, was a time when the climate was comparatively warm. Taber was inclined to correlate the thaw with the Yarmouth interglacial age, chiefly because deep thaw requires a very long time, and the Yarmouth is generally believed to have been the longest of the interglacial ages.[38] Whatever may be the merits of this argument, no evidence of more than one episode of thawing has yet been reported, although Taber clearly stated that the sequence of climatic changes has been much more complicated than the present very incomplete record shows. As study progresses, evidence of freezing and thawing is likely to throw much light on the climatic history of the Pleistocene epoch.

No evidence of an interval of thaw has been reported from the extensively frozen thick alluvial mantle in northern Siberia, although in both Siberia and Alaska slight thaw is now occurring, probably in response to the same climatic amelioration that has been responsible for the world-wide shrinkage of glaciers in modern times. Little help on this problem is likely to come from the greater part of Arctic Canada, because throughout much of that region the mantle is thin and the bedrock lies close to the surface.

RIVER ICE

The Pleistocene alluvial deposits, now terraced, made by the Potomac, James, Tennessee, and other rivers of southeastern United States include cobbles and boulders that are thoroughly striated. The striations can hardly be of glacial origin as the drainage basins of these streams lie wholly outside the glaciated region, and the rock types concerned occur in the bedrocks of the drainage basins. Some of the boulders are very large, and most of the striated pieces are much coarser than the alluvium in which they are inclosed. Although two occurrences are as far south as New Orleans and southeast Texas, most of the others are in a belt about 400 miles wide south of the glaciated region. Striated stones have not been found in the modern alluvium of these rivers, but they are common in the alluvium of large Arctic streams in which the action of river ice during the spring season is powerful.

This collection of facts is taken to mean that the striations were made by river ice, and the stones rafted downstream on ice blocks, at times when the winter climate was more severe than it is today. Probably these times were the glacial ages.[39] It is significant that no striated stones have

[38] Taber 1943, pp. 1539, 1540.
[39] Wentworth 1928; see also Reed 1928.

been found in the preglacial alluvium (Brandywine, "Lafayette") deposited by these same streams.

SAND DUNES AND LOESS

From time to time attempts have been made to draw inferences, some of them far reaching, from the positions and stratification of sand dunes and from the sheets of loess, as to the climates of the times during the Pleistocene when these deposits accumulated. Most of the dunes in question were built during deglaciation, usually to leeward of some rich source of sand such as a mass of outwash or a shore; later both source and dunes became clothed with a covering of plants and dune building came to an end. The evidence is fragmentary, in many instances obscure, and not uncommonly equivocal. Most of it urges conservatism in interpretation. We are hardly justified in drawing many conclusions from this type of evidence.

Dunes

Dunes were built in many places in both North America and Europe during the later part of the Fourth Glacial age. Their presence has been regarded as an indicator of dry climate, and their positions, forms, and stratification have been held to indicate winds different from those prevailing in the dune areas today.

The opinion that dunes record dry climates has been fairly widespread, especially in Germany. Generally combined with the concept is the notion that absence of vegetation in the source area is also a requirement for the accumulation of dunes. This view has been refuted by Cooper,[40] who demonstrated that, although both factors favor dune building, neither is necessary for it. Coastal dunes are forming today in very moist regions such as the coasts of Oregon, Washington, and Alaska, including areas with a well-developed forest cover. Cooper believed that, although dune building can be promoted by dry climate and absence of vegetation, it can be brought about also by lowering of the water table as a result of stream intrenchment. The extensive dune system north of Minneapolis, Minnesota, was in fact developed following intrenchment of the Mississippi River into an outwash mass. Because such intrenchment is a normal event in the history of all outwash masses when the supply of glacial sediment diminishes, and because a very large number of dune areas are located on outwash bodies, intrenchment is probably an important cause of dune building. Certainly the inference that dry climate has been the

[40] Cooper 1935, p. 108.

cause of dune formation is not justified unless specific evidence is brought forward to support it.

In some regions such evidence appears to exist. In northern Arizona longitudinal dunes, tied down by forest vegetation, occur in districts with 12 to 15 inches of rainfall annually. Dunes with this form are being built in that region today only in those districts with less than 10 inches of rainfall. It is concluded that the vegetation-covered dunes indicate a climate drier than that which now prevails.[41]

The positions, forms, and stratification of dunes constitute a problem principally of wind directions. In southern Sweden and Norway, for example, the form and orientation of various groups of dunes dating from late in the Fourth Glacial age indicate that they were built by winds blowing from the northwest and north, presumably from the remnant of the Scandinavian Ice Sheet on the Norwegian-Swedish mountains.[42] On the plains south of the Baltic, however, extensive dunes also dating from late in the Fourth Glacial age appear to have been built by west and northwest winds such as characterize the planetary circulation over this region today.[43]

In North America most dunes dating from the Wisconsin Glacial age appear to be related to winds like those of the present day. At Montowese near New Haven, Connecticut, dunes built of reworked outwash at a time when the margin of the shrinking Wisconsin ice sheet stood more than 25 but probably less than 50 miles distant to the north were built by westerly winds such as prevail today. The same is true of dunes near Hartford, Connecticut.[44] Dunes along the southern shore of Lake Michigan, accumulated during the Cary sub-age, also were built by westerly winds such as the winds prevailing today.[45] On the great outwash plain that grew up contemporaneously with the Bloomington moraine in northwestern Illinois during the Tazewell sub-age, dunes built during or soon after the accumulation of the outwash were made principally by southwesterly winds, which are those prevailing now.

All these results, although undoubtedly fairly accurate, are of questionable value as sources of information regarding former prevailing winds, for dunes can be accumulated not by the prevailing winds of the locality but by "minority" winds.[46] Near Mead, north of Spokane, Washington, a large group of active dunes, growing today, caps the scarps of outwash terraces (a very common mode of occurrence). As

[41] Hack 1941, p. 262.
[42] Högbom 1923; Horner 1927.
[43] Horner 1927, pp. 201–203; see also discussion in Chapter 10.
[44] Flint 1933.
[45] H. T. U. Smith 1940*a*.
[46] Flint 1936, p. 1883.

these dunes lie south of the source of the sand they must have been transported by winds with a northerly component; yet the winds that prevail today are southwest. The southwesterly winds are perhaps stronger as well as more frequent than all other winds, but not having access to a source of sand they have not built dunes. The less frequent northerly winds, sweeping across an abundant supply of sand, have succeeded where the prevailing winds have failed.

A somewhat similar case is that of the Arkansas and Cimarron rivers in southwestern Kansas. Grassed-over dunes, no longer forming, lie in belts 10 to 12 miles wide on the south sides of these rivers. Their distribution and stratification suggest that they were built chiefly by northerly winds although today the predominant movement of sand in those districts is toward the north. It has been suggested that these dunes may have accumulated under the influence of winds of glacial origin.[47] However, the dune areas lie 300 miles southwest of the border of the Nebraskan drift and 500 miles from the border of the Wisconsin drift. It seems very unlikely that the Laurentide Ice Sheet could have produced persistent radial winds in districts so far removed from it.

The widespread occurrence in Bermuda of eolianites (described in Chapter 18), apparently accumulated during the several glacial ages, has been taken to indicate that at those times gales over Bermuda were stronger and more frequent than now.[48] The gales seem probable, but it is questionable whether the eolianites could not have been built by the winds that characterize the island today.

In summary, it seems wise to be conservative in drawing inferences as to prevailing winds from sand dunes, at least until the appearance of stronger evidence than is available at present.

Loess

The character and origin of the loess have already been discussed. At this place we have to deal only with the implications of loess as to the climate of the times when it accumulated. Great stress has been laid on the association of loess, in Europe at least, with a steppe climate.[49] The argument is based partly on the occurrence as fossils in the loess of steppe-inhabiting mammals such as the jerboa and suslik, and partly on the fact that loess has not been found in regions, like Britain and western France, that have a very moist maritime climate.

In a broad sense this twofold argument is probably justified. The

[47] H. T. U. Smith 1940*b*, p. 168.
[48] Bryan and Cady 1934.
[49] Cf Penck 1936*a*; 1936*b*.

greatest development of loess in Europe is in the Ukraine, a region that today is rather dry (albeit perhaps less so during the glacial ages). Similarly the thickest and most extensive loess in North America is in the moderately dry western part of the Central Lowland and eastern part of the Great Plains. In both regions, however, the loess is best developed in the neighborhood of obvious major sources of silt — fine outwash sediments spread along large rivers, notably the Missouri and the Mississippi, the Danube, the Dnepr, and the Don.

Although loess is uncommon in moist regions, it is to be found in them. Loess of Wisconsin date occurs in southern New England in the Connecticut Valley lowland, close to a source of abundant outwash silt, in a region with an annual rainfall of 40 to 44 inches, well distributed throughout the year.[50] It occurs also in the vicinity of Boston, Massachusetts.[51]

Along the Illinois River valley, downstream from the border of the Wisconsin drift, the Wisconsin outwash consists of alternating layers of pink, gray, and buff silt with sand and gravel. The river bluffs have a thick cap of loess, and the loess is similarly colored, indicating that it was derived from the outwash while outwash deposition was in progress.[52] It does not seem necessary to demand a semiarid climate for the accumulation of the loess. All that is required is that the valley floor should have been largely free of vegetation instead of protected by it, as it is today.

Loess is certainly far more abundant where the climate is dry than where it is moist. Beyond this general statement, it is doubtful that detailed climatic significance can be attached to loess as such, for the wind should be capable of deflating silt and clay of outwash and other alluvial origin just as long as these masses of fine sediment were kept bare by the rapidly shifting upbuilding streams. Upbuilding would have continued as long as glaciers were discharging meltwater into the streams, or as long as pluvial climates favored the rapid deposition of nonglacial alluvium in piedmont basins. But, when eventually the streams were reduced to normal discharge, a cover of vegetation would have developed, and the areas offering silt and clay for deflation would have been very much reduced.

There are, then, reasons for the belief that extensive sheets of loess may record no more than the presence of a glacial or somewhat dry climate, or the encroachment of glacier ice into the regional drainage basin.

The concept of loess as an indication of a somewhat dry climate has

[50] Flint 1933, p. 986.
[51] H. T. U. Smith and Fraser 1935.
[52] Leighton 1932.

been applied to China.[53] It is suggested that the loess in northern and east-central China, regions which today are said to be too moist for the accumulation of loess, dates from glacial-age conditions. A southward shift of the climatic zones through about 4 degrees of latitude would bring the cool dry continental climate now characteristic of Inner Mongolia to northern China. Aided by the glacial lowering of sealevel with consequent reduction of rainfall from maritime easterly winds, this shift would have brought to northern China distinctly drier conditions favorable to the accumulation of loess. Presumably these conditions were reversed during the interglacial ages.

FLUCTUATIONS OF LAKES IN NONGLACIATED REGIONS

Relation to Glaciation

The arid or semiarid regions in the interiors of all the continents contain many saline lakes and the beds of many more lakes that are now extinct. It was shown in Chapter 1 that Jamieson and Lartet early recognized the generally synchronous relationship between the expansion of those lakes and the large expansions of glaciers elsewhere on the lands. Fluctuating climates were responsible for both.

It has been argued that the expansion of these lakes was the result chiefly of the pouring into them of meltwater from glaciers. This is doubtless true of some lakes, although it may be questioned whether meltwater was the chief cause of the expansion even of these, but it can not be true of the many lakes that grew in size even though they received no drainage from former glaciers.

The causes of expansion of the lakes in the dry regions of middle and low latitudes appear to have been increased precipitation and decreased evaporation, which characterized the climates of the glacial ages in those regions. The changes came about in this manner.[54] During the glacial ages the temperature gradient between the cold glacier-covered high-latitude regions and the low-latitude belts of calms was steeper than it is today, especially in the northern hemisphere where extensive land areas favored the growth of vast ice sheets. Therefore the exchange of air between the two regions was more active than now. This exchange was expressed in increased turbulence in the atmosphere: increased strength and frequency of the cyclonic storms that characterize the polar front region where cold and warm air masses come into contact with each other.

[53] Cf. Movius 1944, p. 57.

[54] For discussion of this problem see Klute 1928, p. 85; Penck 1913; 1914; Huntington 1914.

With the gradual growth of ice sheets in North America, Siberia, and Europe and with the growth of sea ice in the Arctic Sea and in the North Atlantic, high-pressure conditions were established in the air over them, and the entire belt of eastward-moving cyclonic storms (the "belt of westerlies") was shifted progressively southward. In South America a similar northward shift took place, but it was certainly less pronounced because the area of glacier ice in southern South America was comparatively small.

As the belt of cyclonic storms edged southward, regions — especially in the extra-tropical belts of high pressure — that are dry under the climates of today began to experience increased rainfall. At the same time the increased cloudiness that characterizes the belt of westerlies operated to reduce the evaporation rate, already diminished somewhat by the worldwide cooling that had preceded the growth of the ice sheets. The results were increased stream discharge, expansion of existing lakes without outlets, and the creation of lakes in basins previously dry. In short, a pluvial age came into being. The paths of the cyclonic storms appear to have been pushed southward through a distance of about 15 degrees of latitude.

In North America the most noticeable effects are seen in the Basin-and-Range region (lying mainly in Nevada, Utah, Arizona, and New Mexico) because this region contains a great many closed basins. Localities as far south as Mexico City (lat. 20°) were affected, though the effects were far less pronounced than, for example, those in Nevada and Utah, which lay barely 7 degrees of latitude south of the glacier-covered region.

In the Old World, in like manner, pluvial conditions affected chiefly the Mediterranean lands, northern and central Africa, Asia Minor, central Asia, and northern China. However, the effects were felt right down to the equator itself, where evidence of expanded lakes in East Africa is clear and extensive. This fact is not as surprising as it seems at first. The northern half of Africa is the only large subtropical land mass lying north of the equator. Because of this vast expanse of land it is more favorably situated than any other subtropical region to receive winter-season outbreaks of cold polar air. Even today these outbreaks reach as far south as latitude 15°. During the glacial ages, when climatic zones farther north were displaced as much as 15 degrees toward the equator, cold-air outbreaks should have reached the equator itself. The result is evident. Moist tropical air masses moving eastward from the Atlantic Ocean across the moist Congo basin would have interacted with the cold air, with resulting rainfall. Today, in the warm summer season, the East African lakes receive rainfall in the form of convective precipita-

tion. The cyclonic precipitation outlined above occurred only in winter but should have been at least as great as the summer precipitation of today. Hence in the region of the East African lakes the annual rainfall during the glacial ages should have been greatly in excess of present amounts.

It has been argued that so much atmospheric moisture was expended in building up the great ice sheets that little was left for precipitation on the nonglaciated regions, and hence that the pluvial conditions recorded by the geologic evidence of expanded lakes could not have been contemporaneous with the glacial ages. This argument neglects the important fact that when atmospheric circulation is increased the evaporation of moisture from the sea is augmented (despite reduced worldwide temperature), and this moisture is transported and precipitated more rapidly. Because of this fact it is probable that during the glacial ages the amount of moisture available for precipitation both along the borders of the former ice sheets and in the nonglaciated regions was considerably greater than it is in the same regions today.

This discussion leads to the conclusion that the pluvial ages of expanded lakes were contemporaneous with the glacial ages — that the two reached their maxima at the same times. The evidence in support of this conclusion is both direct geologic evidence of the close association of expanded lakes and glaciers and indirect deductive evidence that the growth of the former ice sheets should have provided the atmospheric mechanism required to produce the pluvial phenomena. The case is not entirely closed, but the probability is of a high order.

North America

The Basin-and-Range region in western North America is marked by 126 closed basins.[55] Of these, 98 record former lakes or expansions of existing lakes, 8 have not yielded unmistakable evidence of former lakes, and the remaining 20 have not been examined in sufficient detail to warrant inference. The 98 basins recording former lakes actually contained only 71 individual lakes, because some lakes flooded more than one basin. The former lakes Bonneville and Lahontan together occupied more than 25 basins.

The distribution of present and former lakes is shown in Figs. 85 and 86; in Table 27 are assembled data on the former water bodies.

In connection with Figs. 85 and 86 Meinzer pointed out that the relative areas of the lakes suggest that the least dry parts of the Basin-and-

[55] Free 1914.

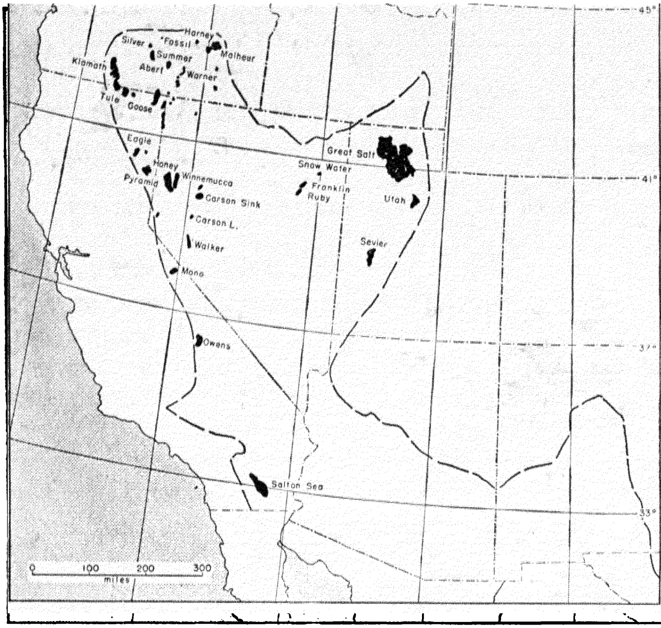


FIG. 85. Existing lakes in the Basin-and-Range region in western United States. (Meinzer.)

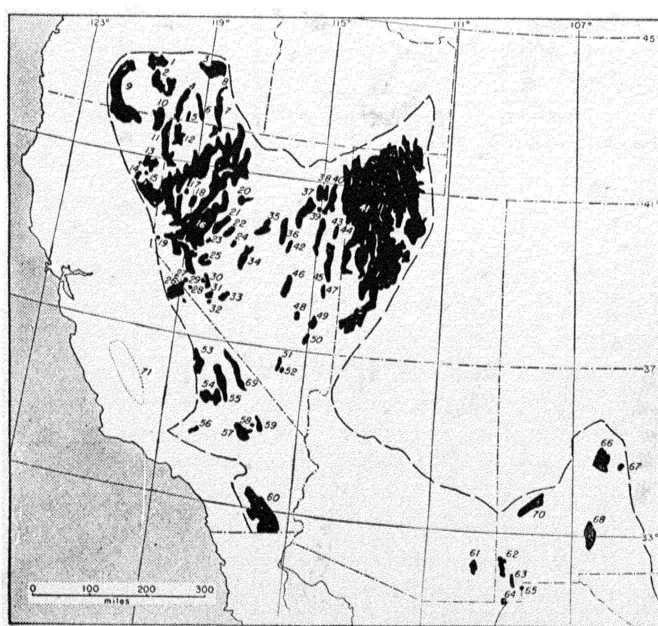


FIG. 86. Pluvial lakes in the Basin-and-Range region in western United States. (Modified after Meinzer.) Compare Fig. 85.

TABLE 27. PRELIMINARY LIST OF PLUVIAL LAKES IN WESTERN UNITED STATES
(Modified after Meinzer)

Map No (Fig. 86)	Name of Basin	Name of Former Lake (if different)	Outlet	No of Strand-lines Recorded	Approximate Maximum Former Depth of Water (feet) above Present Lake or Dry Floor	Selected References
1	Silver, Christmas, Fossil					Waring 1908, p. 30.
2	Abert and Summer	Ls Chewaucan and others		8	250	Waring 1908, p. 51; Allison 1945
3	Harney		Yes			Waring 1909, p. 46.
4	Warner			5-6	200	Waring 1908, p 49; I. C. Russell 1884, pl 83.
5	Guano				*	I C. Russell 1884, pl. 83.
6	Catlow			3	75	Waring 1909, p 64.
7	Alvord			6	100+	Waring 1909, p 70
8	(Unnamed)					I C. Russell 1885, pl 1.
9	Klamath		Yes			Idem
10	Goose		Yes			I. C. Russell 1884, pl. 83
11	Surprise			7	350	I C. Russell 1885, pl. 46.
12	Long					Idem
13	Madeline		Yes			Idem
14	Eagle Lake		Yes			Idem
15	Horse Lake		Yes			Idem
16	(Several basins)	L. Lahontan		Several	886	Idem
17	White sand					Idem
18	Granite Springs					Idem
19	Washoe		Yes			Idem
20	Buffalo Valley		Yes			Idem
21	Dixie			3	150+	Idem
22	Edwards Creek					Idem
23	Fairview					Idem
24	Smith Creek				•	Idem
25	Gabbs					Idem
26	Mono			Many	900	Idem
27	Whisky					Idem
28	Teels Marsh					Idem
29	Excelsior Flat					Idem
30	Soda Spring					Idem
31	Columbus Marsh			3	140	Hicks 1916, fig 1
32	Fish Lake					I. C Russell 1885, pl 46
33	Big Smoky	L. Tonopah		4	50+	Meinzer 1917, p. 57, pl. 1
34	Big Smoky	L Toiyabe			100+	Idem
35	Grass					I C Russell 1885, pl. 1
36	Diamond		Yes			Idem
37	Ruby			13	100+	Sharp 1938, p 317.
38	Independence					I C Russell 1885, pl 1
39	Butte					W O. Clark and Riddell 1920, pl 3
40	Goshute					I C Russell 1885, pl. 1
41	(Several basins)	L. Bonneville	Yes	Several	1050	Gilbert 1890
42	Newark					I C. Russell 1885, pl. 1
43	Steptoe		Yes			W. O Clark and Riddell 1920, p 20.
44	Antelope					Idem
45	Spring					Idem
46	Railroad			11	155	Free 1914, p 36.
47	Duck		Yes			Carpenter 1915, pl 1
48	Coal					Idem
49	Bristol					Idem
50	Delamar					Idem
51	Indian Spring					Idem
52	(Unnamed)		Yes			Meinzer 1922
53	Owens		Yes		250	H S Gale 1914, p 268
54	Indian Wells, Salt Wells, Searles		Yes*	22	600+	Idem
55	Panamint		Yes		930	Idem
56	Antelope					Thompson 1929, p. 303; pl. 19
57	Manix-Afton	L Manix	Yes	3	200-300	Buwalda 1914; Blackwelder and Ellsworth 1936
58	Cronise	Little Mohave L	Yes	2 3	20	Thompson 1929, p. 538.
59	Silver and Soda	L Mohave	Yes	Several	40+	Idem, p 563
60	Salton Sea†			Many	300+	Mendenhall 1909, p 17
61	Sulphur Spring	L Cochise			50+	Meinzer and Kelton 1913, p. 74.
62	Animas			Several		Schwennesen 1918, pp 27, 86
63	Playas			1	40	Idem, p 108
64	San Luis	L Cloverdale				Huntington and others 1914.
65	Hachita					Schwennesen 1918, p. 120.
66	Estancia			Many	150	Meinzer 1911, p. 10.
67	Encino			1	60+	Idem, p 76
68	Tularosa	L Otero		Many	280	Herrick 1904
69	Death	L Manly	?	Several	600	Blackwelder 1933.
70	San Augustin				165	Powers 1939.
71	Tulare		?	2 phases (known from sediments only)		Barbat and Galloway 1934, pp. 492, 496.
72	Long	Long Valley L		Several	250	Mayo 1934.

* Overflow during earlier stage, not during later
† Former lakes not necessarily pluvial.

Range province today are comparable with the driest parts of the province during the pluvial ages.[56] Thus southwestern New Mexico and southeastern Arizona had about as much lake area during the pluvials as southern Oregon has today. The distance between these two areas is 8 to 10 degrees of latitude, and the difference between their mean annual temperatures is about 8° C. Hence there is some basis for the belief that the temperature in the more southerly region was reduced by about this amount. This is of course a rough calculation, which can not take account of the variations in size and depth of basins, increased rainfall, and cloudiness as offsets to lowered temperature, and other variable factors. But it is an interesting method, and Meinzer's results deserve to be weighed carefully with estimates of glacial-age temperature reduction arrived at by other methods.

Lakes Bonneville and Lahontan

Lake Bonneville, by far the largest and in many respects the best known of the former lakes in western North America, occupied a number of coalescent intermont basins in Utah, Nevada, and Idaho. Today the only water bodies in these basins are the highly saline Great Salt, Provo, and Sevier lakes. When at its maximum Lake Bonneville had an area of 19,750 square miles and an extreme depth of about 1050 feet.

The earliest event in the record is a rise of the water surface to a high level (the pre-Bonneville level), where it remained for a long time. Later the water evaporated almost or entirely to dryness, and the basin was dry throughout a long interval. This event was followed by renewed filling to a high level (the Bonneville level) 90 feet above the previous high shoreline, and more than 1000 feet above the present Great Salt Lake. This level was controlled by an outlet (Red Rock Pass) at the north end of the lake, through which the overflowing water spilled out to the Snake River and the Pacific. Erosion of the outlet rapidly reduced the lake level to 625 feet above the present lake. At the new level a hard ledge of bedrock held the lake at a nearly constant level, and the well-developed Provo shoreline was established. Thus from a climatic point of view the Bonneville and Provo strandlines can be considered as a unit, because the fact that the two are distinct is merely a result of erosion of the lake outlet.

Later strong evaporation set in, and the lake shrank with many fluctuations. The most notable is represented by the Stansbury strandline, 330 feet above the present lake.

[56] Meinzer 1922, p. 549.

That the earliest recorded lake stage was preceded by a very long dry time is indicated by the enormous alluvial fans upon which the lake sediments lie.

Gilbert's study[57] of Lake Bonneville was so thorough that during the half century subsequent to it very little has been added. Most of the inferences regarding climatic changes have had to be drawn from the strandlines of the former lake. Little has been learned from the lake-floor deposits because good exposures of them are lacking. What little is known indicates that desiccation intervened between the stages of high water.

The record, however, clearly shows two stages of high water separated by a stage during which the basin was nearly or wholly dry. The earlier high-water stage lasted at least five times as long as the second. At the close of the second high-water stage the process of gradual evaporation was marked by one notable interruption, recorded by the Provo strandline, as well as by a number of minor fluctuations. The whole of this process, including the interruptions, was much shorter than the great desiccation that intervened between the two high-water stages. Gilbert believed that the great desiccation represents an interglacial stage. Upham referred that episode to the Sangamon stage,[58] and, although he later changed his mind, very likely he was right. We may tentatively correlate the Lake Bonneville features as follows:[59]

Evaporation, with fluctuations	Wisconsin stage	Later Wisconsin substages
Stansbury strandline		
Evaporation		
Second high-water stage (Bonneville and Provo strandlines)		Iowan substage
Desiccation	Sangamon stage	
First high-water (pre-Bonneville) stage	Illinoian stage	
Antecedent stage of desiccation	Yarmouth stage (perhaps plus earlier time)	

This correlation agrees with the statements made by Gilbert about relative duration. Presumably there were earlier lake stages whose records have not come to light. Traces of earlier strandlines may have been entirely destroyed, but a complete record should exist in the sequence of lake-bottom sediments buried beneath the present floors of the basins.

A different lake history was suggested by Antevs.[60] According to his scheme the Provo and Stansbury strandlines were reoccupied, so that each strandline was a product of two lakes of different dates. The

[57] Gilbert 1890.

[58] Upham 1914.

[59] Blackwelder (1931, p. 914) suggested a different correlation.

[60] Antevs 1945, p. 26.

inferred sequence of events is therefore somewhat more complex, and may be shown thus:

SEQUENCE IN BONNEVILLE BASIN	SUGGESTED CORRELATION
Desiccation	
Stansbury strandline (reoccupied)	
Provo strandline (reoccupied)	Mankato
Stansbury (low-water) strandline	
Bonneville and Provo strandlines	Iowan
Desiccation	
First high-water stage	An earlier glacial stage

This scheme resembles that suggested above in recognizing two Wisconsin substages plus one pre-Wisconsin stage.

Lake Lahontan,[61] which occupied several basins in Nevada, Oregon, and California, had an area of 8422 square miles (less than half the area of Lake Bonneville) and a maximum depth of more than 500 feet. Pyramid Lake, Winnemucca Lake, the Carson Lakes, and Walker Lake are its principal reduced successors. Its three major strandlines, less well defined than those of Lake Bonneville, appear to belong to a single major rise and fall of the water surface. But the well-exposed sections of lake-bottom deposits, including chemical precipitates, apparently indicate three and perhaps four high-water stages separated by times when the basins contained as little water as they do today or perhaps were entirely dry.[62] Lahontan, like Bonneville, was preceded by a long period of dryness recorded by bulky alluvial fans underlying the lake sediments. In its major outlines, therefore, Lahontan history resembles Bonneville history. But how these stages may correlate with the subdivisions of the Wisconsin Glacial age and with the earlier glacial ages is still undetermined.[63]

Other Former Lakes

The basin of Summer Lake in southern Oregon carries evidence of two high-water episodes separated by a very low-water episode. The implied history has been correlated with the history of Lake Lahontan, the high-water times being referred to the Iowan and Mankato subages respectively.[64]

During the latest glacial age Mono Lake in eastern California attained a depth of nearly 900 feet. At least two high-water stages are recorded, and at least one of these, as already indicated, was contemporaneous with

[61] I. C. Russell 1885; J. C. Jones 1925; 1929; Antevs 1925*a*.

[62] Antevs 1925*a*, p. 101.

[63] Antevs (1945, p. 30) suggested a sequence of events closely similar to those inferred by him in the Bonneville basin.

[64] Allison 1945.

a glaciation. These stages are referred by Blackwelder[65] to the Tahoe (early Wisconsin?) and Tioga (later Wisconsin). In addition there are records of at least two stages in which the water surface was only slightly higher than now.

Farther south, in a region intensely arid today, Owens Lake formerly rose to a level 250 feet higher than now. It overflowed southward, filling the basins of China and Searles lakes. The water in the Searles basin, 600 feet deep, overflowed eastward into the Panamint basin to form a lake 930 feet deep. This in turn apparently spilled eastward into Death Valley, forming a lake 90 miles long and 600 feet deep. There was thus a chain of five lakes. It is possible that the lake in Death Valley drained southeastward to the Colorado River, but the evidence is obscure.

A 104-foot borehole in the floor of Death Valley reveals pairs of layers, each pair consisting of a clay member (deposited in a fresh-water lake) overlain by a salt member (deposited in a saline lake undergoing progressive evaporation). Interpreted on this basis, the section records three lakes of long duration and several transient ones. This unique boring indicates what would probably be found beneath the floors of the majority of basins in the arid west.

The chain of lakes is regarded as possibly earliest Wisconsin in age.[66] In the Searles basin is the fresher strandline of a later lake, not high enough to overflow, which may be later Wisconsin.[67]

In the arid Afton basin, a depression in the lower part of the Mohave River drainage system in southeastern California, three lake stages are evidenced by lake-bottom deposits and by strandlines. The column, with suggested correlations based on extent of erosion of lake beds and degree of decomposition of beach pebbles, is as follows:[68]

Complete desiccation	
Third Lake stage (strandline alt. 1637 ft.)	"Very late Wisconsin"
Complete desiccation	
Second Lake stage (strandline alt. 1795 ft.)	Tioga?
Complete desiccation	
First Lake stage (strandline alt. 1775 ft.)	Tahoe?

The parallel with the Lake Bonneville column is striking.

In the southern part of the San Joaquin Valley, California, the Tulare formation (now deformed) includes fossil-bearing lacustrine beds that record two extensive lake stages separated by conditions of great shrinkage of the lake. The expanded stages have been correlated with glacial

[65] Blackwelder 1931, p. 889.
[66] Blackwelder 1933.
[67] Blackwelder 1941.
[68] Blackwelder and Ellsworth 1936.

stages.[69] Although formerly thought to be later Pleistocene in age, the Tulare formation is now generally regarded as earlier Pleistocene; hence the glacial stages represented may be the Nebraskan and Kansan.

The North American record has been very incompletely studied, but future investigations promise a rich harvest of information regarding Pleistocene climatic changes. The impression created by some writers that the lake evidence supports the concept of a twofold pluvial history is too simple and too rigid to meet the facts. It is far more likely that in the strandlines and lake deposits thus far recorded we have only the later part of the record, and that in order to obtain the full record we shall have to await the sinking of boreholes like that in Death Valley.

Mexico and South America

The evidence from America south of the United States is extremely meager. According to Jaeger[70] evidence of former extensive lakes is abundant throughout the Mexican plateau down to its southernmost part. Details are available only for the region near Mexico City. Here Lake Texcoco (now artificially drained, and within historic time never more than 10 feet deep) rose to a maximum height of 175 feet as shown by lake-floor deposits and by the terraces of tributary streams. The date of the deposits is established as Pleistocene by the occurrence in them of the elephant *Mammuthus* (*Archidiskodon*) *imperator*, but more precise dating has not been attempted.

In the Atacama desert in northwestern Argentina evidence is reported that the existing lakes were formerly considerably expanded, indicating former moister climate.[71]

The basin of Lake Titicaca, at latitude 16° south on the Peru-Bolivia border, is fringed by laminated clay up to more than 300 feet above the present lake. From this sediment a former, much larger lake (Lake Ballivián) is inferred. As the clay is overlain in places by outwash, it antedates at least one glaciation.[71a]

Asia

Aral-Caspian-Black Sea System

The Caspian Sea[72] is the world's largest lake. It is fed by the Volga and Ural rivers, and its surface lies 85 feet below sealevel. Abandoned

[69] Barbat and Galloway 1934, p. 497.
[70] Jaeger 1926.
[71] Walther Penck 1920, p. 251.
[71a] Cf. Moon 1939; Newell 1946.
[72] Davis 1905; Penck 1936*a*.

strandlines are present around its basin up to an altitude of at least 250 feet above the lake or 165 feet above sealevel. They are faint; apparently the water surface fluctuated so rapidly that it did not stand for long at any one position. The distribution of both strandlines and fossil-bearing lake-floor sediments indicates that at one or more times the water rose high enough to become confluent with water in the basin of the Aral Sea nearly 400 miles to the east, and at the same time backed up the Volga at least as far as Kazan. As the lake surface fell, water from the expanded Aral Sea flowed into the Caspian through the Usboi, a remarkable meandering channel now dry, 3500 feet wide and 65 feet deep. Similarly the Caspian appears to have been temporarily confluent through the long Manytch depression (now followed by the Baku-Rostov railway) with the Sea of Azov and the Black Sea. The Black Sea was then a lake, with a surface well above its present surface. It overflowed at the Bosporus into the Mediterranean, whose level was lower than now owing to the large volume of sea water locked up on the lands as glacier ice. Later the level of the Black Sea lake was lowered and the Manytch depression became a river flowing into this lake from the Caspian.

Although it is known that these events occurred, it is not known how many times they were repeated. Grahmann recognized four expansions and three shrinkages of the Black Sea water body.[73] In the Caspian basin he found evidence of at least three lake expansions, separated by at least partial desiccations, and suggested correlations as follows:

Caspian Stages	Suggested Correlation with the European Glacial Succession
Rise to present level	
Shrinkage	Climatic Optimum
Three minor expansions	[Late Fourth Glacial fluctuations?]
Khvalynsk expansion	Fourth Glacial [maximum]
Desiccation	
Khosar expansion	Third Glacial stage
Desiccation	
Baku expansion	Second Glacial stage
Desiccation	

From the stratigraphic sections and abandoned strandlines in the many closed basins around the Aral Sea, Gerasimov inferred a very similar sequence of events,[74] correlated in much the same manner:

[73] Grahmann 1937*a*.
[74] Gerasimov 1930, p. 1088.

Stage	Climate	Suggested Correlation with European Glacial Column
Present	Desert	
Second Interpluvial	Hot; arid	(Climatic Optimum?)
Second Pluvial	Warm; rather moist	Fourth Glacial stage
First Interpluvial	Hot; arid	Third Interglacial stage
First Pluvial	Warm; moist	Third Glacial stage
Lower Quaternary	Hot; moist	Second Glacial stage?

Unlike the lakes of western North America, the Aral-Caspian-Black Sea system of water bodies was augmented by very large volumes of glacial meltwater. The Aral Sea received increments from the glaciers of the Pamir and T'ien mountains in central Asia, via the Oxus River (Amu Darya) and Syr Darya. The Caspian received the vast discharge of the Volga, which drained a large sector of the Scandinavian Ice Sheet. The Black Sea water body received much meltwater from the ice sheet via the Don and the Dnepr, and from the Alps glaciers via the Danube. In addition the Caucasus glaciers contributed to the waters of the Black Sea and Caspian basins. The discharge from these water bodies through the Bosporus into the Mediterranean must have been very large, including, as it did, the drainage from most of central and eastern Europe.

Other Asiatic Lakes

The Dead Sea[75] is an intensely saline lake fed chiefly by the River Jordan. It is now about 47 miles long, and its surface stands at the remarkable altitude of 1300 feet below sealevel. Lake deposits, at least 15 abandoned strandlines, and archeologic evidence show that in Paleolithic time the lake stood 1430 feet above its present level and thereby attained a length of 200 miles. Around its shores Paleolithic habitations flourished in areas too dry today to support a permanent population.

In this region Blanckenhorn was able to distinguish two principal pluvial stages and other changes:

Minor fluctuations of climate.
Dry stage.
Younger pluvial stage (with two substages).
Dry stage.
Older pluvial stage.

In central Turkey several basins without outlet contained more water during the glacial ages than they do now.[76] Lake Burdur (or Buldur)

[75] Blanckenhorn 1912; 1921–1922.
[76] Louis 1938. For Lake Burdur see also Huntington 1914, p. 564.

rose 300 feet and, like Lake Bonneville, flowed out over a spillway. Lake Tuz rose about 250 feet above its present level but had no outlet. Konya basin, south of Lake Tuz, is now dry, but strandlines and fossil-bearing lacustrine sediments show that it formerly contained a lake. Farther east Lake Van rose to 200 feet above its present height. All these lakes are interpreted as products of one or more glacial ages. It is thought that reduced ice-age temperatures alone could have caused the expansions, even without increased precipitation.

In northwestern Iran Lake Urmia, during one or more glacial ages, expanded to nearly twice its present area.[77]

The basin of the two Kara Kul lakes on the Pamir Plateau in the Tadzhik S.S.R. has two sets of abandoned strandlines, a relatively fresh series less than 150 feet above the present lakes and a much older series 150 to 320 feet above the lakes. These suggest two pluvial stages, which were held by Pumpelly to be directly related to glacial stages in the adjacent Trans-Alai Mountains.[78]

Both lake-floor sediments and terraces establish the fact that Lake Issyk Kul in the Kirghiz S.S.R. was greatly expanded at least once, and that this expansion coincided with an expansion of glaciers in the T'ien Shan.[79]

The deposits around the basin of Shor Kul, a salt lake in Sinkiang about 80 miles northeast of Kashgar, yield the following record. The lake expanded to about 350 feet above its present level. Then followed desiccation during which the lake floor was covered with alluvium and in places channeled. Then the lake re-expanded to about 100 feet above its present level. Whether desiccation occurred before the present lake was established is not apparent.[80]

Sairam Nor, a lake in Sinkiang northeast of Alma Ata, was formerly 200 feet higher than now.[81]

Lop Nor, a large swampy lake at the mouth of the Tarim River in central Sinkiang, has at least six strandlines around its basin at altitudes up to 600 feet above its present surface.[82] There is evidence that between expansions the lake contracted to small proportions. Penck thought it likely that the lowest part of the Tarim basin had been the site of a very large pluvial lake.[83]

The Turfan basin in eastern Sinkiang, 200 miles northeast of Lop Nor,

77 Bobek 1937, p. 165.
78 Pumpelly 1905, pp. 138–145.
79 Prinz 1909.
80 Pumpelly 1905, p. 210.
81 D. Carruthers 1914, p. 619.
82 Huntington 1907, p. 356.
83 Penck 1931, p. 13.

has two abandoned strandlines, the higher of which is 200 feet above the basin floor.[84]

In Indian Tibet the Pang-gong Tso and Pangur Tso basins contained contemporaneous lakes during at least two glacial ages, the later generation having been smaller than the earlier. In addition, distinct variations in lake volume have occurred since the climax of the latest glaciation.[85]

AFRICA

The most detailed information on lake fluctuations in Africa comes from the lakes in Kenya and Abyssinia.[86] The many lakes in this region are said to yield a uniform record of climatic changes, as to both intensity and sequence. Because this record lies close to the equator and is very similar to the records of lakes in Asia and North America, it furnishes some support for the belief, expressed earlier, that the climatic changes have been contemporaneous throughout the world.

The evidence can be summed up as follows:

1. Lakes Nakuru, Elmenteita, and Naivasha in Kenya expanded to form a single great lake.
2. Lakes Zwai, Abjata, Langenno, and Shalla in the Abyssinian section of the Great Rift Valley expanded to form a single great lake.
3. Lake Tana on the Abyssinian Plateau was greatly expanded, and southeast of it was formed a far greater lake (Lake Yaya) where now no lake exists.
4. All these lakes alike yield a nearly uniform record of three pluvial stages, as follows:

Postpluvial stage (small-amplitude alternations between moist and dry conditions).
Last Pluvial stage (including five distinct moist substages).
Last Interpluvial stage.
Great Pluvial stage.
(Interpluvial?)
First Pluvial stage.

The two glacial stages on the high mountains of East Africa are correlated with the Last Pluvial and Great Pluvial stages.

From similar evidence around the lakes in Uganda, notably Albert, Victoria, and Kioga, Wayland inferred this sequence of climatic stages:[87]

[84] Huntington 1907, p. 356.
[85] Hutchinson 1939; Huntington 1906.
[86] Nilsson 1931; 1935; 1938; 1940; a dissenting opinion is that of Solomon 1939.
[87] Wayland 1934. The results have been disputed by Solomon (1939)

Uganda Lakes	Suggested Correlation
Postpluvial	Late Fourth Glacial and recent fluctuations
Pluvial 2	Third and Fourth Glacials
Interpluvial (dry)	Second Interglacial
Pluvial 1	First and Second Glacials
Prepluvial (arid)	Pliocene

Although the basis of correlation is not stated and is open to question, the record of two distinct pluvial stages appears to be well established.

From evidence in Kenya, including evidence from some of the lakes examined by Nilsson, Leakey derived the following sequence,[88] which he regarded as closely comparable with the Dead Sea-Jordan Valley column already referred to:

Present conditions	(= Sub-Atlantic wet phase of Europe?)
Nakuran wet phase	(= Climatic Optimum?)
Dry phase	
Makalian wet phase	
Very dry stage	
Gamblian Pluvial stage (2 phases)	
Dry stage, accompanied by faulting	
Kamasian Pluvial stage (2 phases)	

In Egypt Beadnell showed that the now-dry Kharga Depression, 85 miles long, was formerly occupied by a lake more than 300 feet deep.[89] More detailed evidence has been obtained from the Faiyum Depression. The column reported[90] from evidence of former lakes in this basin, and from related evidence, is as follows:

Minor fluctuations, tending toward increasing dryness.
Second Pluvial (less pronounced than First Pluvial).
Interpluvial (desiccation).
First Pluvial.
Dry.

It should be stated that Sandford found in the terraces of the Nile Valley no evidence of climatic fluctuation but only evidence of progressive desiccation beginning at the time of the Acheulian culture stage.[91] He rejected the suggestion that the Nile terraces, necessarily standing above the present river profile, may represent only interglacial stages when the sealevel was high and the climate presumably dry, and that the glacial stages, correlating with low sealevel, are represented by stream trenching, evidence of which is concealed beneath the river. In fact he put forward

[88] Fleure and others 1930, p. 374; Leakey 1930; see also Moreau 1933.
[89] Beadnell 1909, p. 110.
[90] Caton-Thompson and Gardner 1934, pp. 12–18.
[91] Sandford 1936.

the extreme view that the association of pluvial stages with glacial stages has not been established even in a general way.[92]

Australia

In central Australia David noted that the existing lakes were formerly much larger than now.[93] Lake Eyre, for example, had about ten times its present area. This dry region, with predominantly anticyclonic atmospheric conditions, would receive greater rainfall if the belt of Antarctic cyclones, now largely confined to the south coast, were pushed northward toward the equator. This should have occurred during the glacial ages when the Antarctic ice was notably expanded.

Summary

Summarizing the climatic evidence afforded by the lake basins of the dry belts, we may say that, although not conclusive, this evidence points to the alternation of pluvial stages with dry stages. Almost certainly in the Bonneville and Mono lakes the pluvial stages correspond with the glacial stages, and the dry with the interglacial. By analogy this correspondence is true of most if not all of the other former lakes as well. Strong support for this correlation lies in the probability that the growth of ice sheets and sea ice in high latitudes shifted the belt of cyclonic storms equatorward over regions normally dry.

The term *interpluvial* may be used appropriately for the dry stages recorded in the lake basins, not only because of the obvious alternation of these stages with pluvial stages but also because the term conveniently parallels the term interglacial as applied to regions in higher latitudes.

Whether a uniform number of pluvials and interpluvials marked the Pleistocene epoch throughout all low-latitude regions, and how many major fluctuations there were in all, are questions as yet far from solution. The problem is much like the problem of the glacial drift sheets. It has not been reported that as many as four distinct glacial stages are exposed in any single section of drift or at any one locality. The number of sections exposing as many as three drifts is very small indeed, although two drifts exposed at one place are fairly common. The same is true of the lake-basin evidence. The evidence of earlier lake features tends to be destroyed by later erosion. A widespread tendency to lose sight of this fact is evident in attempts to correlate two pluvials with four glacials

[92] Sandford 1935.
[93] David 1932, p. 95.

on no better grounds than the implied belief that the lake evidence noted by the observer must span the entire Pleistocene epoch. So wide a span is not found in most glaciated regions; there appears to be no basis for the belief that it should be found in regions of former lakes.

On the other hand there should exist nearly complete evidence beneath the floors of basins that have been continuously closed. Sections of alternating fresh-water lake deposits, saline precipitates, and in some places also erosion channels and alluvium should be present. Future borings into basin floors can be expected to remove all doubt as to the course of Pleistocene climatic changes in dry regions. In closed basins such deposits must be preserved except for the relatively small amounts that can be removed by the wind during dry stages. The lake deposits are unlike the glacial drifts, which can be extensively removed by interglacial streams and by glaciers during subsequent glacial ages.

ALLUVIUM, STREAM EROSION, AND SOILS

Compared with the interpretation of evidence of former lakes, attempts to reconstruct the sequence of climatic changes in nonglaciated regions by inferences from stream terraces and alluvium meet many difficulties. The regimen of a stream is the result of the complex interaction of many variable factors; a given change in regimen can be brought about in more than one way. It is always difficult to draw climatic inferences from the evidence in one part of a single stream, but to infer that a given climatic change will affect all streams in the same way is disastrous. Much of the thinking about Pleistocene stream history has followed this rigid pattern. Sandford well expressed the necessity for caution in this respect,[94] although implied extension of his criticism to include the evidence of former lakes does not seem as fully justified.

In most attempts to reconstruct climatic changes the prime factor regarded as affecting the regimen of a stream is increase or decrease in rainfall, with resulting fluctuation of runoff and therefore stream discharge. But many other factors enter into the picture, among them (1) topographic setting (whether the stream is in mountains, piedmonts, or plains); (2) absence or presence and character of plant cover; (3) kind of bedrock present, especially in terms of the rate at which rock waste is produced from it; (4) degree to which the condition of the stream approximates the profile of equilibrium; (5) fluctuations of sealevel (if the stream in question is near the sea); (6) fluctuations of glaciers within the drainage basin with resulting changes in the amount of outwash sediment contributed to the stream.

[94] Sandford 1935.

Because these factors and others besides must be considered in the study of every stream, no universal rule of stream behavior with changing climate can be laid down. It is doubtful whether in a moist region an episode of increased rainfall would leave a recognizable record in the valley of every large stream. On the other hand it seems probable that the same episode, in a region normally dry, would be recorded in a major stream by trenching and by increasingly coarse grain size in its bed load, and that a return to a dry climate would result in alluviation with diminishing grain size, provided all other factors, notably the influence of outwash, are neglected. In accord with this view is the opinion that massive fans in arid western North America were built during interpluvial ages and dissected during pluvial ages.[95] But the different opinion is on record that in dry Anatolia massive alluvium accumulated during pluvial ages, owing largely to the rapidity of mechanical weathering at those times, and was dissected during interpluvials despite the smaller stream discharges of those drier ages.[96]

All this does not imply the opinion that every attempt to read climatic changes from stream features is foredoomed to failure. It does imply that studies of stream history must painstakingly consider all the contributing factors before their results are likely to be generally accepted. Many sound and valuable inferences have been drawn from studies of stream terraces and alluvium, and there is no doubt that work on these features, carefully planned and executed, can contribute in no small measure to solving the problem of the Pleistocene climates.

The evidence from soils is less complex and conflicting, and, although very few soils studies with climatic objectives have been undertaken hitherto, the method promises to add much of value to our knowledge of Pleistocene climatic changes.[97]

The basin of Seyistan, where Iran, Afghanistan, and Baluchistan meet, exposes layers of alternating red and yellow alluvium which are believed to record the deposits of moist and dry times respectively. The alluvial deposits are regarded as redeposited soils whose color reflects the climate prevailing at the time they were reworked.[98]

Similarly in Burma De Terra and Movius found that red soils are forming in the moist parts of Burma today whereas the soils in the dry parts of the country are yellowish. By analogy, alternating red and yellow soil layers in Burma are interpreted as recording significant climatic fluctuations.[99] Widespread red-loam soils in northern China

[95] Blackwelder 1931, p. 919.
[96] Pfannenstiel 1940, pp. 418–419.
[97] For a good discussion of details see Bryan and Albritton 1943.
[98] Huntington 1907, p. 363.
[99] De Terra and Movius 1943.

were regarded by J. S. Lee as evidence of a former warm dry climate,[100] although no detailed section has yet been worked out.

From a study of soils and alluvium in the drainage basin of the Vaal River in South Africa, H. B. S. Cooke derived the following climatic sequence:[101]

Recent	Semiarid
Pleistocene	Arid Third Wet stage Semiarid Second Wet stage Semiarid First Wet stage Arid Wet?
Pliocene	Arid?

Some may see in this section a correlation with the Second, Third, and Fourth Glacial ages and the Climatic Optimum, but the evidence hardly justifies such correlation as yet.

From soils and evidence of former lakes Passarge inferred former pluvial conditions in the Karroo Desert and the southern part of the Kalahari Desert in southern Africa.[102]

According to Speight there is evidence in Australia of former climates both moister and drier than the present climate. Moister climates are said to be indicated by bulky deposits of alluvium and stream channels that could not have been cut under existing dry conditions, as well as fossil evidence of both plants and animals.[103]

In the Saharan region of North Africa two moist stages, inferred from evidence furnished by fossil mammals, are separated by a stage of aridity represented by an accumulation of gypsum in the soil.[104] The complex networks of stream channels, not now occupied by water-courses, in parts of the Saharan region have given rise to much speculation concerning their origin. It is believed probable that these streamways were trenched during a pluvial age or pluvial ages when the rainfall on North Africa was greater than it is today. This does not imply that during the pluvials North Africa had no desert country. On the contrary there is reason to believe that large desert tracts persisted throughout the pluvial ages.[105]

[100] Lee 1939, p. 371.
[101] H. B. S. Cooke 1941, pp. 51–52.
[102] Passarge 1904, pp. 648–668.
[103] Speight 1914; 1921.
[104] C. E. P. Brooks 1932, p. 88.
[105] Some of the scattered and fragmentary data on this neglected subject have been assembled in Antevs 1928*a*, pp. 34–38.

The bearing of the fossil record on the Pleistocene climates, and the underlying causes of the climatic changes, are considered in Chapters 23 and 22.

TALUSES

In the arid Basin-and-Range region in western United States taluses have been observed to increase in both number and individual size from south to north and from low altitudes to high, indicating that talus creep is favored by a cold moist climate. Many of the taluses do not now appear to be actively growing; in fact some of them are wasting away. It has been suggested that these accumulations had their active growth during the glacial ages when the climate in the Basin-and-Range region was colder and moister than now.[106]

In the Hopi Country in northeastern Arizona are great blocks of massive sandstone, reaching 1000 feet in individual length, that have slid down over shale. Most of these slides are not contemporary. It has been suggested[107] that they date from a pluvial age or ages when there was greater moisture to lubricate the slides. Dune sand beneath one of these blocks of sandstone may represent a dry interpluvial age.

SUMMARY

The weight of evidence makes it seem probable, though it does not prove, that the glacial ages were contemporaneous throughout the world. If this was true then the interglacial ages also were contemporaneous. During the glacial ages the regional snowline was depressed on a world-wide scale, on the equator as well as in high latitudes, and the amount of depression increased with increasing snowfall and with increasing latitude.

The evidence furnished by frost heaving and solifluction shows that rigorous climatic conditions existed beyond the ice sheets during the glacial ages. On the other hand little can be inferred about Pleistocene climates from either sand dunes or loess.

The expansion and shrinkage of lakes in the dry regions, especially in the extratropical belts of high pressure, apparently took place synchronously with the expansion and shrinkage of the glaciers. The lake record furnishes the most convincing evidence, aside from the drift sheets themselves, of the succession of glacial and interglacial climates. Subsidiary evidence, still very fragmentary, is afforded by alluvium and soils in various parts of the world.

[106] Blackwelder 1935.
[107] Reiche 1937.

Chapter 21

CLIMATES SINCE THE MAXIMUM OF THE FOURTH GLACIAL AGE

THE CLIMATIC OPTIMUM[1]

Botanical Evidence in Europe

We know more about the climates that have prevailed during and since the disappearance of the great ice sheets of Wisconsin time than about the climates of earlier times, because the evidence in peat bogs, lake and marine sediments, and alluvium is abundant, fresh, and comparatively undisturbed. This evidence consists primarily of fossil plants. It is most complete in bogs or swamps in which the accumulation of organic matter has been continuous since these places emerged from beneath the wasting glacier ice. For the most part low temperatures and nonoxidizing bog conditions have preserved this plant material so well that genera are readily identified. When the plants have been identified and the relative proportions of each have been determined for each layer in a deposit, it is easy to reconstruct from this information the general nature of the climate that has characterized the region of the deposit in question at the time each layer was accumulated.

The results are startling to those who had assumed that since the last ice sheets were at their maxima the climate has become progressively warmer and drier. The evidence of the fossil plants and, in addition, several entirely independent lines of evidence establish beyond doubt that the climate (with some fluctuation) reached a maximum of warmth between 6000 and 4000 years ago; since then (again with minor fluctuations) it has become cooler and more moist down to the present time. Apparently as recently as 500 B.C. the climate was still slightly warmer than that of today. The warm, relatively dry interval of 2000 years' duration has been called the *Climatic Optimum*. It is the outstanding fact of so-called postglacial climatic history. This term was devised by Scandinavians, for whom increasing warmth meant amelioration. It is therefore entirely appropriate for northern regions. On the other hand the warmth

[1] See Cooper 1942*a*; 1942*b*; W. B. Wright 1937; Andersson 1910; Gams and Nordhagen 1923; Deevey 1944.

and drought it represents are the reverse of optimum for the drier regions nearer the tropics. Nevertheless the term has come to have a worldwide significance and has been applied, for example, to Asia Minor, for whose people these conditions are not optimum at all.

First recognized in 1885 by Russell[2] from evidence furnished by saline lakes, and inferred independently by Praeger[3] from the fauna of marine deposits in Northern Ireland, this climatic phenomenon has been made clear principally through the work of Scandinavian botanists and geologists, among them Blytt, Nathorst, Gunnar Andersson, Sernander, and von Post. The earlier studies in this field emphasized the physical stratigraphy of peats and fine sediments laid down in lakes, streams, and the sea. As a result there was established the *Blytt-Sernander hypothesis*[4] of the succession of postglacial climates. The principal climatic phases recognized through this study, the Boreal, Atlantic, Sub-Boreal, and Sub-Atlantic, are shown in Table 28. This scheme applies specifically to northwestern Europe, although the basic climatic changes it implies are common to a far wider region.

Through conventional stratigraphic study the relation of the various climates to the succession of water bodies that occupied the Baltic Sea basin (Chapter 15) was established; it is shown in generalized form in Table 28. During the long life of the Baltic Ice Lake, probably some 10,000 years, the climate was cold, as indicated by the remains, at places that stood above the lake, of plants characteristic of the Arctic tundra. The climate became slightly less cold for a time, because birch and pine appeared, only to be superseded by tundra once more before the Baltic Ice Lake was drained. Thereafter, while the short-lived Yoldia Sea existed and the ice sheet had shrunk to very small size, birch and pine reappeared and pine reached its maximum. The time of the Ancylus Lake saw pine give way to hazel as warmth increased. By this time the ice sheet had disappeared entirely. The subsequent Littorina Sea was coeval with the Climatic Optimum. A mixed oak forest developed in large parts of northwestern Europe, but before the Littorina Sea gave way to the modern Baltic deterioration of the climate had already set in, for beech and then pine appeared, and thereafter the oak forest diminished and conditions gradually approached those of the present.

A hypothesis of climatic succession somewhat broader than that of Blytt and Sernander was established by L. von Post as a result of the rapid development of special techniques required for the study of fossil

[2] I. C. Russell 1885, p. 224.
[3] Praeger 1892, p. 212.
[4] Cf. Andersson 1910, p. 197.

pollen. This hypothesis provides a climatic succession consisting of three phases (Table 28) as follows:

3. Decrease of warmth-loving trees and appearance or return of predominant trees of the present day.
2. Culmination of warmth-loving trees.
1. Phase of increasing temperature.

The von Post scheme resembles its predecessor in recognizing a climatic optimum, but it is simpler and more flexible. It fits all parts of the world more easily, because it is not specialized to conditions in northwestern Europe.

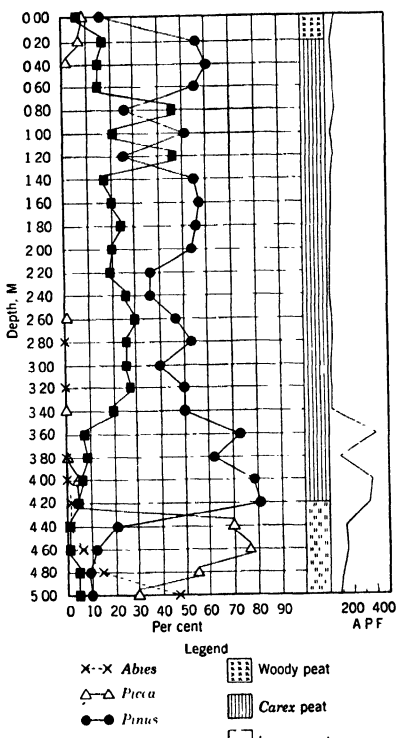


Fig. 87. Typical pollen profile. From a bog near Minneapolis, Minnesota. (Cooper, after Artist.)

The modern system of pollen analysis on which von Post's scheme is based is made possible by the vast quantity of well-preserved fossil pollen in the peat bogs and lake-floor sediments of northern countries. Most of the trees of these regions produce pollen, and it has been established that the relative amounts of the various kinds of pollen now accumulating in a bog or lake are very roughly proportional to the relative abundance of the corresponding kinds of trees that grow in the neighborhood.

Core borings are made in bog peat or lake-floor sediments with a special hand-operated tool. At selected intervals, usually every few inches, the core is treated so as to separate the pollen grains from the remaining cellulose and other material. About 150 pollen grains are counted and identified by genus, and the pollen of each genus is recorded as a percentage of the total number of grains. The percentages of the various genera at each level are then plotted on a diagram. The typical diagram or "profile" represented in Fig. 87 shows, from right to left, the frequency of all pollen grains at each level, the types of peat that constitute the core, and the relative frequencies, at each level, of pollen of the four genera of trees identified in the

TABLE 28. RELATION OF POSTGLACIAL CLIMATES TO STRATIGRAPHIC UNITS AND MAJOR WATER BODIES IN NORTHWESTERN EUROPE
(Greatly modified after H. Movius, *The Irish Stone Age*. See also Movius 1940, Fig. 2.)

*Estimated Chronology (years)**	*Glacial Substages*		*Water Bodies in Baltic Sea Basin*	*Climatic Phases* (Blytt-Sernander)	*Climatic Phases* (von Post)	*Dominant Vegetation*
1946 A.D. 0		Emergence	Modern Baltic	Sub-Atlantic (cool, moist)	Decrease of warmth-loving trees, appearance or return of forest constituents of today	Beech increasing
				Sub-Boreal (warm, dry)		Mixed oak forest diminishing
B.C. 2000		Submergence	Littorina Sea	Atlantic (warm; moist)		Pine increasing; beech, hornbeam appearing
4000					Culmination of warmth-loving trees	Mixed oak forest (oak, elm, lime) dominant
6000		Emergence	Ancylus Lake	Boreal (warm; dry)		Hazel maximum Pine diminishing
				Pre-Boreal (cold)		Pine maximum
8000	Bothnian	Submergence	Yoldia Sea			Birch, pine
			Baltic Ice Lake			Tundra plants
				Arctic (cold)	Temperature generally increasing	
10000 12000	Scanian			Subarctic (less cold)		Birch Pine
14000						
16000	Pomeranian (Danish)			Arctic (cold)		Tundra plants
18000 20000	Brandenburg					

* For limitations on this chronology see Chapter 18.

core, 5 meters in length. The succession of peat types shows that the bog was formerly a lake that gradually became filled with plant matter. The relative frequency profiles show that, in the early stages of filling, the dominant trees were fir (*Abies*) and spruce (*Picea*). These were soon reduced to small numbers and pine (*Pinus*) became dominant. Much later oak (*Quercus*), which had been present all along, took the lead. But it soon diminished, to be followed successively by pine and spruce. This record indicates a cold climate that gradually ameliorated, reaching a climax of warmth (presumably the Climatic Optimum) in the oak phase. Subsequently the climate became cooler and moister, reaching present conditions at the top of the diagram.

Although the evidence of a climatic optimum has been established in greatest detail in Scandinavia and Finland, it occurs throughout Europe. Abundant indications in the Alps suggest that during the Optimum the regional snowline was at least 1000 feet higher than it is today. The result must have been the destruction of scores if not hundreds of small glaciers in places where glaciers exist now. It follows that a large number of the present glaciers in the Alps are not survivors of the last glacial maximum, as was formerly universally believed, but are glaciers newly created within roughly the last 4000 years.[5] During the Optimum average summer temperatures were about 1.5° C higher and the summer growing season about 15 days longer than at present.[6]

The Climatic Optimum has been recognized in marine deposits, not only in Northern Ireland as already indicated, but also in Norway. In the sediments laid down in the Littorina Sea and now exposed on the sides of Oslo Fjord, the fossil mollusks show that water temperatures fluctuated from time to time, reaching a maximum of 2.5° C higher than at present.[7]

The fact that the largest glacier in Iceland, Vatnajökull, is delivering birch logs and shells of marine mollusks at the termini of its outlet glaciers suggests that the outlets were recently shorter than now, and that they subsequently re-expanded over birch forests and marine deposits. This suggestion is best explained in terms of a climatic optimum.

Evidence from Pollen in North America[8]

The pollen studies made in North America, although less far advanced than those made in northwestern Europe, nevertheless combine to sug-

[5] von Klebelsberg 1935, p. 573.

[6] A good summary of the European climates, and an extensive bibliography, are given by Gams and Nordhagen 1923.

[7] Brøgger 1900–1901.

[8] See Cooper 1942*b*; Deevey 1939; 1944; G. D. Fuller 1939; Raup 1937; Sears 1935.

gest so nearly the same climatic succession that similarity of postglacial climates on both sides of the Atlantic is highly probable. Curiously enough no North American profile has yet yielded a record of tundra plants like those of the Arctic phase in Europe. The American profiles begin with spruce and fir. This fact means either that no belt of tundra plants separated the margin of the shrinking Laurentide Ice Sheet from the encroaching spruce-fir forest or (as seems more likely) that no deposit thus far sampled extends back to the actual uncovering of the locality by the ice sheet. The question remains unanswered.

The succession of vegetational types, inferred climates, and inferred correlation with the Blytt-Sernander scheme is set forth in Table 29. It is interesting that climatic cooling since the Optimum appears to have been more effective in eastern Canada and Wisconsin, the two most northerly areas, than in the areas farther south, as indicated by the return of spruce.

Although the Climatic Optimum in northeastern North America was both warmer and drier than the present climate, the same phase in central North America may have been distinguished primarily by increased dryness without great increase in temperature.

In localities in Connecticut, Illinois, and Wisconsin the early (Pre-Boreal?) spruce zone shows two maxima (not represented in Table 29) separated by a spruce-fir flora that indicates a less cold climate. As each of these localities in Illinois and Wisconsin lies just south of the end moraine of one of the Wisconsin substages, it is believed that the return of spruce reflects the same downward fluctuation of temperature that brought about a re-expansion of the ice sheet. In Connecticut the glacial evidence is obscure, but by analogy a re-expansion of the ice sheet there is suspected. This example illustrates one way in which pollen studies may lead to a refinement of purely glacial investigations.

Evidence from Marine Invertebrates in America

At Boston, Massachusetts, marine sediments deposited during the disappearance of the last great ice sheets consist of two zones distinctive because of their content of fossil invertebrates. The lower and older zone is characterized by a fauna whose living representatives inhabit the relatively warm waters between Cape Cod and Cape Hatteras but do not now live as far north as Boston. The higher and younger zone contains a fauna indicative of colder water. The conclusion is drawn[9] that the change is the direct result of a climatic chilling that began some thou-

[9] Shimer 1918, p. 458.

TABLE 29 POSTGLACIAL VEGETATIONAL AND CLIMATIC SUCCESSION IN EASTERN NORTH AMERICA BASED ON POLLEN STUDIES (DEEVEY 1939, 1944)

Eastern Canada (Auer)	*Connecticut (Deevey)*		*New York (McCulloch)*	*Ohio (Sears)*	*Wisconsin (Voss)*	*Climate*	*Suggested Correlation with Blytt-Sernander Phases in Europe*
	Coast	*Inland*					
Spruce	Oak-chestnut	Spruce-hemlock	Oak-hemlock	Oak-beech	Spruce	Cool; moist	Sub-Atlantic
Oak maximum	Oak-hickory		Oak maximum	Oak-hickory	Oak-mixed deciduous	Warm, dry	Sub-Boreal
Hemlock-oak	Oak-hemlock		Oak-hemlock	Oak-beech		Warm, moist	Atlantic
Pine	Pine		?	Pine	Pine	Warm; dry	Boreal
Spruce-fir	Spruce maximum		Spruce-fir	Spruce-fir	Spruce-fir	Cool	Pre-Boreal
	Spruce-fir						
Not represented in North American pollen diagrams							Arctic Subarctic Arctic

sands of years ago. There can be little doubt that the change marks the end of the Climatic Optimum.

EVIDENCE FROM SALINE LAKES AND OTHER FEATURES IN AMERICA[10]

In the eighties the detailed study of Glacial Lake Lahontan suggested to I. C. Russell[11] that Winnemucca, Pyramid, and Walker Lakes did not contain nearly as large a concentration of salts as they should contain if, as had been supposed, they were the result of the evaporation of a former body of the vast size of Lake Lahontan. He therefore inferred that that great lake had evaporated to dryness and that the existing lakes had been created anew at a much more recent time. It was shown later that Pyramid and Winnemucca lakes probably had been freshened as a result of salts being carried out of their basins by a stream that drained them into a lower desert basin, and that Russell's inference therefore applied only to Walker Lake. However, in his basic concept Russell laid the foundation of important evidence of the Climatic Optimum.

Confirmation of Russell's opinion came much later, not, as it happens, from the Lake Lahontan region but from Owens Lake, much farther south, at the eastern base of the Sierra Nevada. Comparison of the concentration of salt in Owens Lake with the amount annually discharged into the lake by its tributary, Owens River, resulted in an estimate that the lake is about 4000 years old. Hence it is inferred that the large and deep Owens Lake of the Wisconsin Glacial age evaporated to dryness during the Climatic Optimum, and the bed of salt residual from its evaporation was buried beneath a covering of alluvium and wind-blown sediment. With the reappearance of a cooler and moister climate some 4000 years ago the present lake appeared. Because the salt left by its predecessor was buried, the new lake was fresh, and it has not yet had time to attain a high concentration of salt.

This interpretation has an interesting bearing on the history of the near-by glaciers of the Sierra Nevada. Owens Lake receives its water almost entirely from Sierran snows, for direct precipitation in the area between the Sierran summits and the lake is insignificant. The same snows that maintain the lake also nourish the Sierran glaciers. Hence it is probable that at the time when Owens Lake was dry the glaciers likewise were nonexistent.[12]

Essentially the same history applies to the lakes within the basin of

[10] See an excellent summary in Matthes 1942, pp. 204–214.
[11] I. C. Russell 1885, p. 224.
[12] Matthes 1942, p. 212.

the former Lake Lahontan, and to Abert Lake and Summer Lake in southern Oregon. There is good reason to believe, indeed, that all the lakes of the Basin-and-Range region have had an essentially similar history.

A climatic optimum is suggested likewise by the delta of Bear River, a stream of glacial meltwater that flows into the Portland Canal at Stewart, on the Alaska-British Columbia boundary. Because three accurate surveys were made during the period 1909–1927, it has been possible to compute the volume of the delta with fair accuracy, as well as the approximate volume of sediment contributed to the delta each year by the Bear River. At the present rate of sedimentation the delta then is estimated to be only 3600 years old, although there is reason to believe that the area was deglaciated and that the sealevel assumed its present position relative to the land much longer ago than that figure would indicate. Accordingly it is suggested that the glaciers that provide the Bear River with its load of silt today did not come into existence until 3600 years or more ago (after the Climatic Optimum). Most of the time during and prior to the Optimum the site of the delta had lain beneath glacier ice and had therefore received no delta deposits.[13]

A climatic optimum is suggested also by the discovery of buried forests now being exposed above timberline on Mt. Hood in Oregon. It appears likely that these forests flourished when the climate was milder, and when the glaciers of Mt. Hood were much reduced in size or had disappeared altogether.[14]

Similarly in the Glacier Bay and Yakutat Bay districts of Alaska, Cooper found a forest, laid bare by recent glacier shrinkage, that had been deeply buried by an expansion that culminated within historic time, at least 45 miles beyond the buried forest. He considered that the forest was established at least 2000 years ago and might be a record of the Climatic Optimum.[15]

Again in the Stikine-Taku region in northern British Columbia pieces of large trees are appearing in the termini of two glaciers.[16] This fact can only mean that these glaciers were formerly less extensive than now, and that they have expanded through forest-covered valleys. The forests may be interglacial, but it is more likely that they date from the Climatic Optimum.

In the Monarch Valley on the west slope of the Front Range in Colorado, the glaciers are believed to have disappeared entirely after their

[13] Hanson 1934.
[14] Matthes 1932, p. 284.
[15] Cooper 1942*a*.
[16] Kerr 1936, p. 691.

latest expansion in Wisconsin time.[17] During this time the glacially eroded rock basins below the high cirques became completely filled with peat, under a climate apparently at least as warm as the present one. Subsequently (but probably prior to 1000 years ago) the glaciers were reborn or re-expanded. This sequence of events strongly suggests the Climatic Optimum.

Many of the small glaciers that occupy cirques in the Sierra Nevada are fronted by one or more very fresh end moraines that stand in sharp contrast with the much more mature-looking moraines a little farther down the valleys. The whole sequence of moraines suggests, not a continuous series made by gradually dwindling glaciers, but two series separated by a considerable lapse of time.[18] This lapse of time may be the time of the Climatic Optimum.

An ingenious study of a group of soils in west Texas led to the recognition of alternating moist and dry climates within the past 25,000 years, although the climatic changes have not yet been dated closely enough to justify the identification of any of them with events elsewhere.[19]

The relations of dunes to fossil-bearing alluvial units, in the Navajo Country, Arizona, suggest that the climate there was formerly drier than now. This dry time may have been the Climatic Optimum.[19a]

One strong suggestion of the Optimum comes from an unusual source —a study of the temperatures in a copper mine on the Keweenaw Peninsula in northern Michigan. The present mean annual temperature at the surface of the ground in this district is 44.3° F In the tens of thousands of years of the Wisconsin age during which the region was blanketed by the Laurentide Ice Sheet, heat was abstracted from the ground through a considerable depth, and the surface temperature was reduced to that of the base of the ice, about 32° F. When the ice sheet disappeared higher surface temperatures returned, and heat penetrated the ground, in time warming the rocks to considerable depths. Temperature readings made in the mine at short intervals down to 6000 feet are plotted as a curve. By deducing what form the curve should have on various assumptions as to the time since deglaciation and as to the changes in the surface temperature, and comparing the result of each assumption with the curve obtained by actual readings, it is found that the best fit is obtained on a single assumption. This assumption is that the district was deglaciated 20,000 to 30,000 years ago and that after an

[17] Ives 1938, p. 1064.
[18] Matthes 1939, p. 520.
[19] Bryan and Albritton 1943.
[19a] Hack 1941, p. 263.

interval the climate became warmer than now and subsequently reached approximately its present condition.[20]

Many islands in the Pacific Ocean are reported to show a persistent strandline at +5 feet. This is said to be an emerged strandline, distinct from any bench cut by storm waves with the sea at its present level.[21] This strandline shows by its state of preservation that it is very young. If truly separable from the present-day strandline, it may constitute a record of the Climatic Optimum, when glacier ice on the lands was somewhat smaller in volume than it is at present. On the other hand the end of the Optimum in the region of Great Britain has been correlated[22] with a slowing down of the rate of rise of sealevel rather than with an actual change in the direction of sealevel fluctuation. Presently available data are insufficient to establish the relation of the Climatic Optimum to the level of the sea.

As in the Alps, so in North America the Climatic Optimum apparently witnessed the disappearance of many glaciers as the snowline rose. Matthes suggested that all the present glaciers of the Sierra Nevada, nearly all those in the Rocky Mountains within the United States, and all the lesser glaciers in the Cascade and Olympic mountains date back only through the time since the Optimum, a time which he appropriately termed a "little ice age."[23]

Evidence from Other Continents and from the Sea Floor

Of the climates of the last few tens of thousands of years little is known outside Europe and North America. C. E. P. Brooks pointed out that the water of Lake Chad in French Equatorial Africa is nearly fresh, although if it dated from the last glacial age it should be saline. He inferred that the lake must have been reconstituted in comparatively recent time, following an episode of desiccation.[24]

There is some evidence on the fringes of the great "sand seas" of central and western Australia that a moister climate is now destroying dunes built during an earlier episode of great drought.[24a]

Archeologic investigations have established the fact that in prehistoric time the Sahara had a climate much less dry than now and was densely populated. Archeologic evidence suggests that desiccation began in the

[20] Hotchkiss and Ingersoll 1934.
[21] Cf. H. T. Stearns 1945*b*.
[22] Godwin 1943.
[23] Matthes 1939, p. 520.
[24] C. E. P. Brooks 1932, p. 91.
[24a] Browne 1945, p. xvii.

neighborhood of 8000 to 6000 B.C. The lower figure can be readily taken as the beginning of the Climatic Optimum.

From excavations at the buried former city of Ashkabad (in the Turkmen S.S.R.) Pumpelly concluded that the city grew up about 8000 B.C. and that cereal crops were then cultivated in the surrounding region. At that period both the Sahara and the trans-Caspian region enjoyed a climate somewhat like that of central Europe today, because the reduced ice sheet still lay over Fennoscandia and the belt of cyclonic storms lay well south of its present position. Both regions later, apparently at the time of the Optimum, became too dry for agriculture, and Ashkabad as well as the Saharan communities fell into ruin. The first great urban communities were established while the region of greatest modern industrial progress was still partly covered by the Scandinavian Ice Sheet.[25]

In New Zealand the pollen researches of Cranwell and von Post[26] have established a threefold climatic sequence since the last deglaciation: (1) cold and dry; (2) warm and moist; (3) the present rather cold climate. The middle member is undoubtedly the Climatic Optimum. A piece of supporting evidence may be mentioned also. Exposed in alluvium in the Canterbury Plains is a buried forest consisting of a kind of tree (*Podocarpus totara*) characteristic of a warmer climate than the present one.[27]

On both coasts of the Bering Strait and on Melville Island the remains of trees are found as much as 500 miles beyond the present timber line, and conversely in western Siberia the taiga, the subarctic forest, is encroaching on the steppe land to the south. The former may record the Optimum while the latter indicates the cooler post-Optimum climate.

Invertebrate fossils in postglacial marine sediments at several localities near the Strait of Magellan indicate a climate postdating the last extensive South American glaciation but warmer than the present climate.[28]

A thorough review of the evidence of climatic fluctuations since 2500 B.C. led C. E. P. Brooks[29] to infer two dry times and three wet times during the 4500 years covered. For the most part the data were inadequate to furnish proof of contemporaneous fluctuations, but the dry phase ending about 2000 B.C. he thought so pronounced as to suggest "worldwide or cosmic causes."

[25] C. E. P. Brooks 1922, p. 163.
[26] Cranwell and von Post 1936.
[27] Speight 1916, p. 361.
[28] Halle 1910.
[29] C. E. P. Brooks 1931.

Finally the North Atlantic sea-floor sediments have a contribution to make to this subject. The coring and analysis of these sediments, described in Chapter 20, yielded this fact. In the four cores in which the uppermost "nonglacial" sediments are thickest this layer contains pelagic Foraminifera of a definitely colder-water type than those in the layer immediately beneath. It is inferred that during the middle part of "postglacial" time the surface water in that part of the North Atlantic was somewhat warmer than the water in the same area today.[30]

MORE RECENT CLIMATIC FLUCTUATIONS

Data gathered by the International Commission of Snow and of Glaciers have shown that, wherever information has been obtainable, glaciers reached a maximum in the nineteenth century and since that time have been shrinking at a very rapid rate. Three hundred years earlier, in the Alps and Iceland at least, the glaciers were no more extensive than they are today, and still earlier fluctuations are indicated by the fascinating story of the earliest Greenland colonies which were established during a time of mild climate in the tenth century and came to grief during a colder period, probably in the fifteenth.[31] The reliability of the evidence varies, but there is no doubt whatever that glaciers throughout large parts of the world, within recent historic time at least, have waxed and waned with remarkable broad uniformity.[32] The uniformity is broadly true of both polar hemispheres at all altitudes. The Greenland Ice Sheet, the Antarctic Ice Sheet, the glaciers of middle latitudes, and those on or close to the equator both in East Africa[33] and in the Andes of South America[34] alike are being reduced in size.[35]

Probably the worldwide net shrinkage of glaciers (with some fluctuation) during the past hundred years is related to two facts, the apparent rise of sealevel during the same period by about 2.5 inches,[36] and the known rise of temperature during the same period. Kincer demonstrated that mean annual temperatures have been rising at a rate varying

[30] Bradley and others 1940, p. vii.
[31] Huntington and Visher 1922, p. 105.
[32] Cf. Matthes 1942, p. 190.
[33] P. C. Spink, *unpublished.*
[34] Notestein *in* Cabot 1939; Broggi 1943.
[35] There are of course local differences that do not affect the principle. A good example is found in southern coastal Alaska. There glaciers with low-level snowfields are shrinking while those with high-level snowfields are holding their own or even expanding. It is believed that the present worldwide increase in temperature is raising the level of maximum snowfall, thereby reducing snowfall on the lower snowfields but at the same time adding to the snowfall on the higher snowfields (D. B. Lawrence, *unpublished*).
[36] Gutenberg 1941, p. 730.

between 0.5° C and 2.2° C per century.[37] The increase is evident not only in the northern hemisphere but in the southern as well. This temperature rise is very likely the main cause of the glacier shrinkage, and the apparent sealevel rise its direct result. It is likely that the condition of glaciers throughout a broad region at any time is a good indicator of the general trend of climate in that region. Finally the apparently worldwide increase in temperature, shrinkage of glaciers, and rise of sealevel during the last few score years increases the probability that the glacial ages and sub-ages likewise have been coeval throughout the world.

[37] Kincer 1933.

Chapter 22

CAUSES OF THE CLIMATIC FLUCTUATIONS[1]

BASIC DATA

The causes of the cold glacial-age climates have challenged the ingenuity of geologists since before the glacial theory became widely accepted. A great deal of thought has been given to the problem by geologists, astronomers, physicists, and climatologists, a dozen or more hypotheses have been offered in explanation, and scores of papers have been written in advocacy of them. Some of the explanations formerly advanced are now definitely out of the running, and nearly all of them find serious objections of one kind or another. Today we are in possession of facts enough to enable us to narrow the field to a small choice and to perceive the general nature of the probable true explanation. Unfortunately, however, there is enough uncertainty in some of our basic data to prevent our being sure. With this preliminary word of caution we may look at the data of which we are reasonably sure, as set forth in the foregoing chapters, and test several competing hypotheses against them. These data are as follows:

1. There were at least four glacial ages and three interglacial ages. The Fourth Glacial age was itself marked by distinct climatic fluctuations including minor fluctuations within historic time.
2. The later fluctuations were (and are) contemporaneous throughout the world; the earlier ones are inferred, by analogy, to have been likewise worldwide, though this is not certain.
3. The fluctuations are not periodic.
4. Their maximum amplitude in terms of mean temperatures is probably of the order of 10° C.
5. The cold and warm climates, respectively, were accompanied by consistent lowering and rising of the regional snowline, amounting to at least 3000–4000 feet in middle and low latitudes. Closely related is the fact that all the areas now occupied by glaciers were formerly more extensively glacier-covered.
6. The climatic belts of the world preserved their general positions relative to each other throughout the Pleistocene epoch and, for that

[1] Good general references include Jeffries 1924, pp. 262–267; Albrecht Penck 1936*b*; 1938; Arldt 1918–1920; Sayles 1922; Speight 1939, p. 63; Hann 1903, pp. 375–403.

matter, throughout the Cenozoic era. The successive glaciers in both North America and Europe repeatedly covered nearly the same areas, and their borders are broadly parallel.

7. Only in the Permian (230 million years ago), Lower Cambrian (about 500 million years ago), and Upper pre-Cambrian (about 700 to 800 million years ago) does the geologic record thus far reveal glacial conditions perhaps comparable in extent with those of the Pleistocene.

HYPOTHESES OF CLIMATIC CAUSES

With this array of basic data we can dispose of some of the hypotheses of the glacial climates. Taken in their entirety the hypotheses can be arranged largely within five main groups:

1. *Topographic hypotheses,* based on changes in the areas and altitudes of the continents and of localized highlands.
2. *Atmospheric and oceanic hypotheses,* based on variations in the amounts of volcanic dust and carbon dioxide in the atmosphere, variations in the salinity of the sea, in the rate of precipitation of moisture on the Earth's surface, and in the strengths and positions of ocean currents.
3. *Geophysical hypotheses,* based on variations in the relative positions of the continents, in the distribution of the continents relative to the Earth's axis, and in the Earth's internal heat.
4. *Planetary hypotheses,* based on periodic changes in the motions of the Earth as a whole, which affect the distribution of solar heat to various parts of the Earth.
5. *Cosmic hypotheses,* based on variations in the absolute amount of solar heat received by the Earth, and on the passage of the Solar System through nebulosities.

HYPOTHESES NO LONGER TENABLE

Several respectable hypotheses should be mentioned for the sake of the record and then dismissed as impossible:

The early, virtually supernatural idea that the ice ages were caused by the passage of the Solar System through "cold regions of space."

The hypothesis, attributed to Humphreys,[2] that loading the Earth's atmosphere with volcanic or cosmic dust would lower the surface temperature of the Earth. The average diameter of the dust particles is greater than the wave length of solar radiation but is only a small fraction of the wave length of terrestrial radiation. The idea is that incoming

[2] Humphreys 1913.

solar heat is baffled out while outgoing reflected heat is passed through. However, astronomers are not entirely in agreement as to what the effect on the Earth's surface temperatures would be. Further, it is extremely improbable that adequate amounts of dust could be held by the atmosphere during the very long periods required to build extensive glaciers. Finally, if dust were the chief cause of glaciation we should find widespread glaciation in the Ordovician, the Devonian, and the Miocene strata, which, although rich in volcanic ash, are free of glacial features.

The hypothesis put forward more than once, that the Solar System has repeatedly passed through nebulosities or cosmic dust clouds, which cut off from the Earth a part of the Sun's heat. It has been shown, however, that the observed nebulosities are thousands of times too discrete to produce a noticeable effect on the Earth's climate. Further, there is reason to believe that such a cloud, even if suitably dense, would raise the mean temperature at the Earth's surface instead of lowering it.[3]

The hypothesis, attributed to T. C. Chamberlin and to Arrhenius, that fluctuations in the amount of carbon dioxide and water vapor held in the atmosphere would affect the absorption of solar radiation by the atmosphere is quantitatively inadequate.

The hypothesis, of which the first may have been that of Whitney,[4] attributing glaciation to increased precipitation without decrease of temperature.

The group of hypotheses that advocate displacement of all or parts of the Earth's crust relative to the Earth's axis, thus bringing different regions into the positions of the poles at various times. Of these the most widely known is the hypothesis of continental drift attributed, in this connection, to Wegener. Aside from the dubious value of the geologic evidence of such displacements, these hypotheses assume, incorrectly, that true polar conditions are the optimum for glaciation, whereas, as we have seen, cold-temperate regions with abundant precipitation are likewise favorable. The consistent maintenance of the relative positions of the climatic zones throughout the Cenozoic,[5] the consistent fluctuations of the regional snowline, always proportional to the present snowline, and the close parallelism of the drift borders of the several glacial ages in North America and Europe render these hypotheses, as Leverett once put it, "purely fantastic."

The group of hypotheses appealing to general and local uplifts of land areas, and to the appearance at strategic points of lands that had not previously existed, as direct causes of glaciation. Such views were

[3] Riives 1941.
[4] Whitney 1882.
[5] Chaney 1940.

put forward as early as 1856 by Dana[6] and were later upheld by Lyell,[7] Upham,[8] and Spencer.[9] As working hypotheses they may have been well enough in the days when only a single Pleistocene glaciation was recognized. But the establishment of the succession of glacial and interglacial ages rendered these views untenable, despite the fact that the repeated uplift and subsidence of Scandinavia were invoked as late as the 1920's[10] in explanation of the repeated growth and decay of the Scandinavian Ice Sheet. Though we condemn these ideas we must admit in fairness that the general concept of uplift has merit and is in fact a highly important contributing cause of glaciation. Also, as far as crustal movement may have played a part in shifting the paths of major ocean currents, uplift may have had an indirect influence on climate. it fails only when applied rigidly as a *direct* cause. Of this more hereafter.

THE HEAT-DISTRIBUTION HYPOTHESIS

The foregoing ideas are outmoded and outdated and are no longer seriously entertained. On the other hand two quite different concepts, both of them current, enjoy great popularity among their adherents. Both are ingenious and rather complex. In all other respects they differ profoundly from each other. One is the heat-distribution hypothesis; the other is the Simpson concept. The basis of the former seems to have been suggested by J. Adhémar in 1832, but the idea first received detailed attention in 1875, when it was elaborated by Croll.[11] Croll's argument ran thus: Disturbances of the Earth as a planet by the Moon and the Sun cause a periodic shift in the position of perihelion. The shift has a period, the "precessional period," of a little more than 21,000 years. The shift affects the *distribution* of solar heat received by the Earth, though it does not affect the *total amount* received. Briefly, the result will be that any given point at a high latitude in the northern hemisphere will be affected as follows: During a fraction of the precessional period it will have relatively long cold winters followed by relatively short hot summers. During the succeeding equal length of time this same point will have slightly shorter winters than before, each with a little more heat per hour, followed by slightly longer summers than before, each with a little less heat per hour. Croll believed that snow accumulation, resulting in the growth of glaciers, would be favored by the period of longer winters and that snow wastage would be reduced by the accompanying

[6] Dana 1856.
[7] Lyell 1875.
[8] Upham 1893.
[9] Spencer 1898.
[10] Wilhelm Ramsay 1924; C. E. P. Brooks 1928, p. 302.
[11] Croll 1875.

shorter summers, despite the greater summer heat. The theory requires that the postulated glacial ages alternate between the northern and southern hemispheres and that the duration of each glacial age be limited to no more than half of the precessional period.

Croll's ingenious concept[12] can not be accepted because the geologic evidence shows that the glacial and interglacial ages have been far longer than can be permitted by the theory, and because the evidence we have, though incomplete, disfavors alternation of climatic changes between the hemispheres. Furthermore, Hann showed that the theory is not valid quantitatively in that the calculated temperature variations are far too small to bring about the results envisaged by Croll.[13]

Croll's idea was supported by Geikie, Ball, and other Europeans and was opposed from the outset by most American geologists, including Gilbert, Chamberlin, Upham, and Le Conte. It has long since been universally discarded. Nevertheless the basic concept of changes in heat distribution as a result of periodic planetary changes persisted, and fifty years after the first publication of Croll's attempt the concept was being newly elaborated and gained many new adherents, virtually all of them Europeans. American students in general were not impressed with the improved versions of the heat-distribution concept, any more than their predecessors of an earlier generation had fallen in with Croll's original version.

The modernized theory is highly ingenious, and is unquestionably an improvement over the original. It recognizes the effects of the precessional period of about 21,000 years essentially as they were put forward by Croll. But in addition it appeals to two other planetary movements which likewise influence the distribution of heat on the Earth's surface. One is the regular variation in the eccentricity of the Earth's orbit, with a period of about 91,800 years. The other is the regular variation in the angle between the Earth's axis and the plane of the Earth's orbit, the period of this motion being about 40,000 years. The results of each of these changes, in terms of the heat received at any given latitude on the Earth's surface at 1000-year intervals, are calculated mathematically. The results are combined and are plotted as a curve, which is carried backward to any desired date. Because the periods of the three component movements differ, the resulting curve is nonperiodic. It shows irregularly spaced maxima and minima of heat which are taken by the advocates of the theory to represent the interglacial and glacial ages respectively. However, the fluctuations in heat reception are regional; the total amount of heat received by the Earth as a whole remains constant.

[12] A good summary in simple terms is given by W. B. Wright (1937, pp. 316–331).
[13] Hann 1903.

Various calculations have been made, notably by Spitaler,[14] Soergel,[15] Köppen and Wegener,[16] and Milankovitch.[17] In the last, which was first made in 1920 and is widely quoted (see Fig. 74), four pairs or groups of temperature minima, calculated for the summer season only, are regarded as representing four principal glacial ages, all of them together having occurred within the past 600,000 years.[18]

The amplitudes of the temperature fluctuations, calculated, we must repeat, for the summer only, are about equal to the differences in mean summer temperature that exist between points separated by 10 to 15 degrees of latitude.

According to this sweeping deduction the First, Second, and Third glacial ages each had two conspicuous and subequal maxima, and the Fourth had three. Furthermore, the interglacial separating the Second and Third glacials was three to four times as long as the other two interglacials. And the whole of the recognizable Pleistocene sequence is crowded into little more than 600,000 years.[19]

This skilfully constructed scheme avoids two serious objections to the Croll idea: the heat fluctuations are not periodic, and they do not alternate rigidly between the hemispheres. On the other hand, neither do they coincide. Two of the basic factors produce exact alternations of heat minima between the hemispheres. The third, the variation in the eccentricity of the orbit, produces minima in both hemispheres at the same time. The combined effect of all three factors is not to produce heat minima in the southern hemisphere coincidentally with heat maxima in the northern, but to offset the heat minima at irregular time intervals between the hemispheres. Temperature fluctuations in the equatorial region would occur, but would be minimal, and some would involve increases, while those at higher latitudes involved decreases.

Seven objections[20] to this general scheme and to the variations upon it thus far published may be mentioned.

1. The scheme fails to consider any factor other than solar radiation, despite the proved close relationship between glaciation and the distribution of land and sea.[21]

[14] Spitaler 1921.

[15] Soergel 1937.

[16] Köppen and Wegener 1924.

[17] Milankovitch 1930; 1938.

[18] According to Spitaler's schedule the Pleistocene temperature fluctuations fall within a period of 1,334,400 years. This disagreement with the Milankovitch scheme weakens the reliance that can be placed on either system.

[19] According to Spitaler, 1,334,400 years.

[20] The most comprehensive criticisms are in Penck 1938 and G. C. Simpson 1940.

[21] C. E. P. Brooks 1928, pp. 117–119.

2. According to Simpson[22] the calculated temperature changes are too small to have brought about the glaciations, and he put forward grounds for the belief that the actual temperature changes attributable to the three basic factors are much smaller even than those calculated by Milankovitch.

3. The scheme demands that the maxima of the deduced glacial ages be not simultaneous in the two polar hemispheres; yet there is no geologic evidence that they were. Indeed most glaciers today are rapidly shrinking simultaneously in the northern hemisphere, in the southern hemisphere, and on the high mountains of the equator itself. Furthermore the temperature rise during the last hundred years has affected both polar hemispheres simultaneously.

4. It requires slightly increased temperatures in the equatorial region during the glacial ages. Yet in the equatorial Andes and in the mountains of equatorial East Africa the regional snowline was greatly depressed at these times. In fact between the glacial and interglacial ages it fluctuated through more than 3000 feet, a figure only slightly smaller than that for the Alps.

Putting items 3 and 4 in a different way, lowering of the regional snowline simultaneously in both hemispheres can not be explained by any regional or even hemisphere-wide diminution of a heat supply that is constant for the Earth as a whole. On the contrary it demands a reduction in the heat received by the surface of the entire Earth.

5. The nine heat minima deduced from the Milankovitch curve do not agree with the geologic evidence, which, though admittedly fragmentary, records a maximum of only four glacial ages.

6. The overall time, about 600,000 years, is far less than the time that the geologic evidence of interglacial soil decomposition seems to demand. If the curve is extended farther back in time, still more heat minima appear, making it embarrassingly difficult to correlate a particular minimum with a particular glacial age.

7. In view of the fact that the periods that constitute the basic factors in the construction of the curve are not precisely known, there is serious question whether the computations, though probably generally reliable for later parts of the time scale, are accurate enough to be valid for constructing a curve for a time some hundreds of thousands of years in the past.[23]

These seven objections, taken together, constitute a formidable barrier to acceptance of the heat-distribution hypothesis, even in its modern form. Yet a few geologists and geographers in Britain and many in

[22] G. C. Simpson 1940.

[23] Cf. Brouwer, *quoted in* Schulman 1938.

Germany have accepted it not only in principle but even in detail.[24] It is difficult to escape the inference that supposed coincidences between the curve and the facts of Pleistocene stratigraphy have obscured, in the minds of proponents of the scheme, the doubtful factors that lie back of the curve itself. The modern version of the heat-distribution hypothesis has attracted almost no adherents in North America, and no attempt at detailed North American comparisons has been made. Perhaps this is because American glacial geologists have been more skeptical of the scheme and of its basis than have many Europeans, although Antevs[25] used the scheme in tentatively dating North American events. On the other hand we must admit that the nature of the geologic evidence in any part of the world is necessarily far from precise, and that close comparison with any mathematical curve is therefore very difficult to make.

Despite the serious objections to this hypothesis it is not implied that the effects of the several heat-distribution factors are negligible. Undoubtedly these factors have influenced climates to some extent. But their effects would seem to have been superposed, as modifications, upon those of some more potent cause of climatic fluctuation.

HYPOTHESES OF FLUCTUATION OF SOLAR RADIATION

The view that the Pleistocene climatic changes have resulted from variations in the *distribution* over the Earth's surface of a constant quantity of solar heat has now been reviewed. Distinct from this is the view that the changes are caused by fluctuation of the ***absolute amount*** of heat received by the Earth from the Sun.

The Earth's surface receives heat from the Sun at the rate of 1.94 calories per minute per square centimeter. This is the ***solar constant.*** Strictly speaking this value is not a constant; it is known to undergo fluctuations of small amount. That fluctuations in the radiant energy of the Sun might be related to the changing Pleistocene climates was perceived as long ago as 1891, a time when Croll's hypothesis was receiving wide and favorable comment in Europe. C. H. Hitchcock suggested that "the sun's heat may have been variable, being considerably diminished during the glacial epochs."[26] In 1895 Dubois published a well-reasoned monograph attempting to show that variations in solar heat were specifically

[24] It should be noted at this point that Penck, although originally an adherent of Croll's hypothesis, changed his position when it became established that the regional snowline had fluctuated widely in the equatorial region. His latest view was that worldwide temperature reductions had a cosmic origin (Penck 1938).

[25] Antevs 1938.

[26] C. H. Hitchcock 1891, p. 237.

responsible for the glacial climates.[27] Not until 1904, however, was it established by actual observation that current fluctuation of solar radiation is a fact. Regular measurements made at the Smithsonian Institution since 1918 have recorded fluctuations in the solar "constant" amounting to about 3 per cent of its average value. "It is quite possible, but not capable of proof as yet, that larger solar changes have occurred in the past and may occur in the future."[28] On the assumption that larger changes do in fact occur, the hypotheses that follow hold that terrestrial climatic fluctuations take place in accordance with them.[29]

Huntington's Hypothesis

Huntington's version of this basic idea laid emphasis, not on direct temperature changes on the Earth, but on a supposed connection between solar energy and the incidence of cyclonic storms in the Earth's atmosphere.[30] It postulated a periodic or near-periodic recurrence of sunspots, associated maximum atmospheric "storminess" with sunspot maxima, and extended the apparent fluctuations of "storminess" within historic time back into geologic time as an explanation of the glacial and interglacial ages. Huntington suggested that sunspot "cycles" are essentially glacial and interglacial ages in miniature. During sunspot maxima, he thought, the belts of cyclonic storms would migrate poleward and the resulting precipitation, taking place in higher latitudes, would include increased proportions of snow which in turn would build glaciers large and small. The hypothesis thus appealed primarily to changes in the amount and character of precipitation rather than to worldwide changes of temperature.

Apparently it is established by observation that storms are most common when sunspots are most numerous, perhaps because at such times horizontal temperature gradients are increased and greater turbulence results. However, whether sunspots recur cyclically or, regardless of cycles, produce long-term storminess sufficient to bring about the results claimed is very doubtful. Aside from this quantitative objection, Huntington's hypothesis also encountered a profound qualitative objection. The apparently worldwide lowering of the regional snowline and the chilling of the sea during the glacial ages appear to require world-

[27] Eugène Dubois 1895.

[28] Abbot 1936, p. 108.

[29] This was the view of Arctowski (1908, p. 6), who, discussing climatic changes, remarked: "Le soleil, auquel nous devons la vie, est peut-être la cause de cette variation, et toute cette histoire des glaciers de la terre n'est peut-être qu'une marque (si profondément ressentie sur notre globe) de quelques-unes de ses pulsations."

[30] Huntington 1914; Huntington and Visher 1922.

wide reduction of temperature. The best the hypothesis could offer in this connection was mere regional lowering of temperature brought about by proximity to the large ice sheets once they had formed — a distinctly limited and secondary effect. Thus, despite the fact that it avoided some of the difficulties encountered by the heat-distribution hypothesis, Huntington's proposal, although stimulating and suggestive in its emphasis on the nourishment of ice sheets by snowfall from cyclonic storms, did not succeed as an explanation of the Pleistocene climates.

Simpson's Hypothesis

In contrast with Huntington's scheme the hypothesis advanced by Sir George Simpson appeals to the effect of fluctuations of the Sun's output of energy on the Earth's temperature, rather than on the storminess of the Earth's atmosphere.[31] Like a hypothesis advanced a little earlier by Grunsky,[32] Simpson's scheme explains glaciation in the first instance through increase of warmth, but differs in most other respects. It is almost wholly deductive. Beginning with the assumption that solar radiation experiences large fluctuations, it deduces the consequent effects on the Earth's atmosphere. The argument runs thus:

Increased solar radiation raises temperatures at the Earth's surface, more at the equator than at the poles. As a result the rate of circulation of the atmosphere is increased. The higher temperatures and stronger winds bring about greater evaporation, which in turn causes increased cloudiness and increased precipitation. The cloudiness reduces climatic extremes and gives climates generally a more maritime character. In relatively low latitudes the resulting conditions are essentially pluvial, and each time of increased solar radiation is marked by a pluvial age. Conversely radiation minima correspond with interpluvial ages.

In high latitudes, however, increased radiation increases both temperature and precipitation. At first snowfall increases with increasing total precipitation, the annual accumulation of snow increases to a maximum, glaciers form, and an ice age reaches its climax. In time, however, as the temperature continues to rise, the proportion of snowfall to total precipitation decreases and melting and evaporation increase. When melting and evaporation come to exceed snowfall, the total annual accumulation of snow reaches zero, the glaciers disappear, and the ice age is at an end. There follows a relatively warm wet interglacial whose peak occurs at the peak of the solar radiation curve. Thereafter events occur in reverse. Radiation diminishes, temperature falls, precipi-

[31] G. C. Simpson 1934; 1940.
[32] Grunsky 1927–1928.

tation decreases, snowfall commences, and, when it exceeds melting and evaporation, snow begins to accumulate once more. Glaciers form, and another ice age comes into existence. But as snowfall continues to diminish with diminishing precipitation, the glaciers dwindle and disappear and a cold dry interglacial follows, its peak coinciding with the low point on the solar radiation curve.

In summary (Fig. 88), two assumed cycles of solar radiation, each consisting of an increase followed by a decrease, should result, in low latitudes, in two pluvial ages separated by an interpluvial. In high latitudes the same two cycles should yield two pairs of two glacial ages, each pair separated by a warm wet interglacial and the two pairs separated by a cold dry interglacial. It is important to note that according

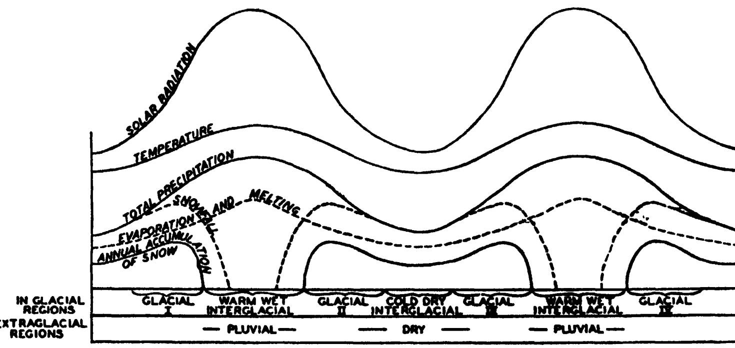


FIG. 88. Effect of two cycles of fluctuation of solar radiation on temperature, precipitation, snow accumulation, glaciation, and the spacing of pluvial and dry times according to Simpson's hypothesis. The curves are qualitative only. (Modified after Simpson.)

to this scheme the pluvial ages and the glacial ages do not coincide in time. The diagrams used to illustrate the scheme are, as Simpson admits, rigid and purely qualitative. Evidently they were designed primarily to attempt to show that the four known glacials and the three known interglacials could be fitted into the deduced consequences of two cycles of solar radiation.

Simpson's concept has two outstanding advantages over the heat-distribution hypothesis. It does not involve rigid limitations to predetermined time intervals, and it does not demand that any climatic changes occur at different times in the northern and southern hemispheres. Furthermore it harmonizes with the view cited in Chapter 4 that the central mass of the Antarctic Ice Sheet could have been more expanded than now only under a climate that was warmer rather than colder than the present one.

On the other hand the scheme encounters formidable difficulties. It

demands noncoincidence of the pluvial and glacial ages; yet the evidence at Mono Lake and Lake Bonneville indicates that the two phenomena were approximately coincident. It requires that for four glacials there be only two pluvials; yet we seem to have evidence, in both hemispheres, of more than two pluvials. It demands two contrasting types of interglacial climates; yet we have no geologic evidence that more than one type existed. In fact the attempt to apply the deductive scheme in detail to the geologic evidence fails in many serious respects.[33] The concept, ingenious as it manifestly is, deserves to be kept in mind and checked against significant new geologic data as they become available. But it can hardly be considered a satisfactory explanation of the facts we now have.

SOLAR-TOPOGRAPHIC HYPOTHESIS

General Statement

The hypothesis offered here in competition with those outlined on the preceding pages is believed to meet the accumulated array of facts more satisfactorily than its predecessors. It is given a name—*solar-topographic hypothesis*—only because in any discussion of the relative merits of the various ideas, many of them complex, the use of a name for each saves space and avoids misunderstanding. The name *solar-topographic* is selected because it conveys the two essential elements in the idea: (1) fluctuation of solar radiation as the cause of worldwide temperature changes, and (2) the presence of highlands as the prime factor determining the accumulation of snow and the distribution of glaciers. No originality is claimed for the basic concepts involved. The concept of former significant fluctuation of solar energy dates back, as stated earlier in this discussion, at least to Hitchcock in 1891 and has been favorably considered by several students of the problem since that time. The close genetic relationship between glaciers and high lands was clearly recognized by Dana as early as 1856 and was undoubtedly in the mind of Agassiz before him. To be sure, later recognition of the wide extent of former ice sheets over comparatively low lands, particularly in North America, tended to obscure for a time the essential importance of the adjacent highlands as the places where the glaciers originated. But the mountains and plateaus without whose presence those ice sheets would never have been built are now beginning to appear in their true significance as the starting places of the ice sheets.

To repeat, then, the solar-topographic hypothesis, named purely for convenience in discussion, is not a new hypothesis. It is a combination

[33] See the discussion in G. C. Simpson 1934, pp. 471–478.

of two ideas put forward long ago, but with extensions and modifications of both. Although Simpson's hypothesis is wholly deductive (Simpson himself explains clearly why it must be so[34]) the solar-topographic concept is largely inductive. It is built on certain observational data which, if not facts, at least possess a very high degree of probability.

Basic Data

These basic data can be stated concisely:

1. During the glacial ages the regional snowline was lowered roughly parallel with itself.
2. The Pleistocene climatic changes affected and are continuing to affect the whole world at the same time,[35] and the pluvial climates coincided with them.
3. The climatic zones in the northern hemisphere maintained their relative positions throughout the Pleistocene epoch.
4. Glaciers are genetically related to mountains and other highland areas.
5. Solar energy is observed to fluctuate at present through a small range.

From 1 we infer that lowered temperature, not increased precipitation, started each glacial age.

From 2 we infer that the temperature was lowered in both polar hemispheres simultaneously, that the cause was probably extraterrestrial, and that the heat-distribution hypothesis fails to explain it.

From 3 we infer that movement of the poles or of continents has played no part in the major climatic fluctuations of the Pleistocene.

From 4 we infer that the glaciers formed initially in areas of high land; be it noted, however, that in extreme polar areas the relative heights of some highlands were not great.

From 5 we can infer only that small-amplitude fluctuation is a fact. This tells us nothing about larger-amplitude fluctuation. All we can say at present is that larger fluctuation is at least possible. It seems unavoidable that the basic cause of temperature changes lies outside the Earth, and solar fluctuation meets the facts better than any other extraterrestrial cause. Yet it must be remembered that change in solar energy greater than that actually observed today is no more than an assumption.

These inferences, taken together, lead to the hypothesis that fluctuations of solar energy are responsible for the major climatic changes of

[34] G. C. Simpson 1934, p. 428.

[35] The Antarctic Ice Sheet is a possible exception, in that the apparent response of this glacier to climatic change seems anomalous and is not clearly understood.

the Pleistocene, that the most conspicuous response to lowered temperatures was the formation and expansion of glaciers in highlands, chiefly in high latitudes, and that important secondary climatic effects resulted from the expansion of some of the glaciers into extensive ice sheets.

THE TOPOGRAPHIC FACTOR

The stage was set by the development of high and mountainous continents. High continents set with newly folded or newly uplifted mountain ranges were the culmination of successive and repeated uplifts in widely scattered parts of the world. Some of these uplifts had their origins as early as Oligocene time; others are traceable to Miocene and Pliocene movements some of which persisted strongly into the Pleistocene.

Despite the fact that references are scattered and the data have never been fully assembled, the worldwide distribution of these movements is striking. In North America late Pliocene or Pleistocene movements involving elevation of thousands of feet are recorded in Alaska[36] and in the Coast Ranges of southern California.[37] In Greenland similar movements are referred to the "Tertiary,"[38] but there is reason to believe that they took place during the late "Tertiary." Conspicuous uplifts of what had been lowlands in Labrador and eastern Quebec occurred "between middle and late Pliocene."[39]

In Europe the Scandinavian Mountains were created from areas of very moderate relief and altitude in "late Tertiary" time.[40] Iceland underwent great faulting movements, with vertical components of more than 6000 feet, as late as the Pleistocene.[41] The Alps were conspicuously uplifted in Pleistocene and late pre-Pleistocene time.[42]

In Asia there was great early Pleistocene uplift in Turkestan,[43] the Pamirs,[44] the Caucasus, and central Asia generally.[45] Most of the vast uplift of the Himalayas is ascribed to the "latest Tertiary" and Pleistocene.[46]

[36] Capps 1931; Taber 1943, p. 1542.
[37] H. R. Gale 1931, pp. 71–72; Bailey 1943.
[38] Koch 1935, p. 149.
[39] H. C. Cooke 1929, p. 119.
[40] Ahlmann 1919, p. 141.
[41] Pjeturss, *cited in* Ahlmann and Thorarinsson 1937, p. 165.
[42] Heim 1919.
[43] Pumpelly 1905.
[44] Popov 1932, p. 45.
[45] Gromov 1945, pp. 509–510.
[46] De Terra 1934; De Terra and Paterson 1939.

In South America the Peruvian Andes rose at least 5000 feet in post-Pliocene time.[47] In Australia the great Kosciusko Plateau uplift occurred at the close of the Pliocene,[48] and in New Zealand there was strong late Pliocene and early Pleistocene uplift (the Kaikoura Orogeny).[48a] In addition to these tectonic movements many of the high volcanic cones around the Pacific border, in western and central Asia and in eastern Africa, are believed to have been built up to their present great heights during the Pliocene and the Pleistocene. Altogether the evidence of highland-making in late Cenozoic time fully supports the concept that glaciation was at least in part a result of these movements. It is noteworthy that among the highlands named are those believed to have been the places of origin of the Scandinavian and Laurentide ice sheets, the two largest in the northern hemisphere.

In addition to these major movements, general uplifts postdating one or more of the early glaciations and amounting to hundreds of feet vertically affected parts of the Rocky Mountains region, the Sierra Nevada, the Alps, the Caucasus, and certain ranges in central Asia. In other ranges, notably the Himalayas, uplifts postdating at least some glaciation were far greater.

The cumulative result of both gradual and successive uplifts throughout the entire second half of the Cenozoic era was an increase in the average height of the continents from an estimated value of less than 1000 feet to their present height of 2500 feet.

This rise in itself should have lowered surface temperatures by an average of at least 3° C, and by considerably more at the higher altitudes. In addition, mountains impede the atmospheric transfer of heat from the equator toward the poles. They dry the air that passes over them, and the dried air cools the sea surface by evaporation. The cloud formation induced by mountains reflects solar radiation and thus aids terrestrial cooling.[49] Hence, in both direct and indirect ways, the rise of highlands increases the likelihood of glaciation.

The creation of new highlands could not, however, have been the sole cause of the glacial climates, else we should have evidence of only one glacial age extending from the beginning of the Pleistocene through the present time. The great shifts of the regional snowline upon the flanks of stable or nearly stable highlands demand a quite different kind of cause. But if the mountains had not been present and if the continents

[47] Berry 1930; Moon 1939.
[48] Browne 1945, p. xiv.
[48a] Gage 1945, pp. 153, 154, 156.
[49] These effects are mentioned by C. E. P. Brooks (1936).

had not been high, no extraterrestrial cause yet suggested in any hypothesis would have been adequate to bring about glaciation. Temperatures would have fallen only slightly, and except possibly in the vicinity of the poles the regional snowline would not have been brought low enough to intersect any part of the surface of the land. Maritime climates would have prevailed, as apparently they prevailed in Cretaceous time and in Ordovician time — to name two extreme examples.

Only at times of extremely widespread building of high mountains should altitudes have been great enough to permit the growth of really extensive glaciers. It is in the rocks of the only such times, which the geologic record shows to have been comparable with the Cenozoic mountain building in height and worldwide extent, that we find clear evidence of widespread glaciation: the Permian, Lower Cambrian, and Upper pre-Cambrian. This earlier record is too dim to permit bringing it farther into the argument. Glaciers at other times of mountain making there may have been — there is at least some evidence of them in the earliest Cenozoic — but unless they had grown to ice-sheet size they would have been confined to the highlands in which they originated, and in those regions of steep slopes and ultra-rapid erosion the evidence of their former presence would soon have been destroyed by erosion.

In summary, fluctuation of solar energy was able to cause extensive glaciation only at those times when the upheaval of highlands reached up, as it were, to meet the falling temperature part way.

In this requirement of high lands as a prerequisite for extensive glaciation seems to lie the key to the comparative rarity of widespread glaciation throughout the annals of geologic history. Only when uplifts were unusually high and widely distributed, *especially in the regions traversed by the belts of westerly winds,* could extraterrestrial heat fluctuation succeed in reducing temperatures enough to bring about the building of great glaciers. According to Chaney,[50] the record of Cenozoic fossil floras is one of diminishing temperature throughout the later Cenozoic. It is also one of diminished rainfall on the lowlands. Presumably it was the distribution of precipitation that was changing, becoming heavier on the windward flanks of the growing highlands as it diminishes on the intervening lower lands. Thus the stage was set for glaciation in terms of precipitation as well as in terms of temperature. Heavy concentrations of rainfall on the flanks of highlands were ready to be converted into snowfall and thence into glaciers by the simple process of worldwide reduction of temperature.

[50] Chaney 1940, p. 486.

The Solar Factor

This reduction of temperature, as already stated, is assumed to have been the direct result of reduction in the rate of emission of radiant energy by the Sun. The fluctuation in the mean annual temperature at the Earth's surface that is recorded by the geologic evidence amounts to about 10° to 11° C. How much variation in solar energy would be required to bring this about? The quantitative relation of solar fluctuation to Earth temperature has been discussed in some detail by Simpson.[51] Extrapolation on Simpson's very helpful chart suggests that a variation in the Earth's mean temperature from 8° C less than now during the glacial ages to, say, 3° more than now during the interglacials would require a fluctuation of solar energy of about 10 per cent on either side of its mean, or 20 per cent in all.[52] However, so many variable factors are involved that no such calculation can be expected to be accurate. Possible climatic results of current fluctuation have been mentioned by Sverdrup,[53] and Abbot[54] found that 0.5 per cent change in the solar "constant" may result, at the Earth's surface, in a temperature change of 2.5° to 4° C.

Growth of the Glaciers

As the temperature reduction occurred, glaciers formed on highlands according to their altitude and to the precipitation on them.[55] In most of these highlands the snowfall was slight enough, or more commonly summer-season wastage was great enough, so that the glaciers formed there never grew so large that they extended beyond the highlands. In a few favored regions, however, low summer temperatures coupled with snowfall that was moderate or even abundant resulted in spreading of the glaciers beyond the highlands themselves to evolve ultimately into ice sheets. The Scandinavian, Siberian, and Laurentide ice sheets

[51] G. C. Simpson 1934, pp. 464–468.

[52] Simpson's hypothesis led him to a requirement of 40 per cent — double the above figure — to satisfy the demands of the pluvial climates as he conceived them. Although the faunal evidence is somewhat conflicting, his concept of the pluvials seemed to demand a far higher temperature and far more precipitation and cloud amount than the geologic evidence warrants.

[53] Sverdrup 1940.

[54] Abbot 1936, p. 111.

[55] It is possible that on very high and very favorably situated mountains glaciers existed before the time now commonly thought of as marking the beginning of the Pleistocene epoch, just as earlier in geologic history they may have existed on high mountains at times other than those generally regarded as glacial. The record of such glaciers is not likely to have been preserved, and it is probable that their former existence will always remain conjectural.

originated in such highlands and spread outward in the directions in which spreading was possible. The first spread east and south, the second formed from the coalescence of radially spreading glaciers from several distinct highlands, and the last, also with more than one highland of origin, spread mainly west and south.

Setting aside, for greater simplicity, the inference discussed in Chapter 17 that the Siberian Ice Sheet spread over a region west of the Ural Mountains from which it was later displaced by the more vigorous Scandinavian ice, we can say that it appears probable that these three great ice sheets, together covering more than eight million square miles when at their maxima, were essentially contemporaneous. After they had attained large size and had begun to spread into lower latitudes, increased summer temperatures and therefore increased rates of wastage began to take their toll and the rate of advance of the glacier margins must have become more and more slow. It is likely, in fact, that toward the climax of each glacial age the regimen of each of these vast ice bodies had reached a condition approaching equilibrium, in which nourishment, wastage, and discharge were about equally balanced. To put it in another way, the ice sheets had spread southward about as far as they could under the existing conditions; a very substantial further decrease in regional temperature would have been required to bring about much additional expansion.

This, seemingly, is the principal explanation of the rather close accordance of the drift borders of the successive glacial ages. Viewed broadly these borders are rather strikingly similar, and those pertaining to glaciations earlier than the Fourth are not too far from being congruent. The corresponding ice sheets were apparently hovering near equilibrium each time the temperature reversed and climatic conditions leading toward an interglacial age set in. During the Fourth glaciation the change in climate set in somewhat sooner, before the ice sheets had reached their former extent. The Scandinavian and Laurentide ice sheets, expanding under similar climatic conditions, failed by similar margins to reach the limits set by their predecessors during earlier glacial ages. The Siberian Ice Sheet, on the other hand, was controlled by a drier climate with warmer summers and in consequence made a conspicuously poorer showing during the Fourth Glacial age than its larger contemporaries. Its separate major parts simply did not grow sufficiently to coalesce and form a single continuous ice body.

From all this we infer that it is not necessarily true that the earlier glacial ages were of about equal lengths. Given a minimum length sufficient for their expansion, the ice sheets would not have spread farther no matter how much longer the low temperatures endured. As

far as we can read the evidence the Earth's heat supply diminished and then increased in at least four major cycles. The clear record of the fourth cycle involves several subsidiary fluctuations of decreasing amount. By analogy we may well suppose that the earlier cycles also were marked by fluctuations the stratigraphic evidence of which has not yet come to light.

Auxiliary Factors

No discussion of the glacial climates would be complete without some reference to the influence of topographic changes on ocean currents. It has been held, for example, that submergence of extensive parts of Central America would permit part of the Gulf Stream to flow into the Pacific, and would thereby reduce the amount of heat now transported by the Gulf Stream to northwestern Europe. Aside from the fact that the faunal evidence records a land connection between the Americas throughout at least a substantial part of the Pleistocene epoch, this mechanism alone is wholly inadequate to explain the glacial climates. That it would have exerted an auxiliary regional influence, if it ever occurred, is not denied.

Another example that has been cited is the possible former rise of the submarine ridge that connects Greenland and Iceland with The Faeroes and Britain. No part of this ridge today lies more than 1500 feet below sealevel. It is held that emergence of this ridge would have barred the warm water of the Gulf Stream from the Greenland or Norwegian Sea and the Arctic Sea, thus cooling the Norway-Spitsbergen region. This, too, would have exerted a regional influence, perhaps locally pronounced,[56] though as a general cause of glaciation it is powerless.

Still other auxiliary factors are more likely than the two just mentioned to have played a part in shaping the glacial climates regionally. These would have begun to operate after the general reduction of temperature had begun and glaciers had formed on the highlands most susceptible to them. The result would have been to promote the growth of the glaciers and to reduce temperatures in the ice-covered regions still further. The descent of the zone of maximum snowfall (Chapter 4) hastened glacier expansion. The reflection of solar heat by the intensely white snow- and ice-covered surfaces, about four times that by snow-free surfaces, materially lowered regional temperatures. The development of ice sheets in the northern hemisphere pushed the atmospheric polar front southward and fended off warm tropical air masses, thereby still further contributing to the cooling effect. In the Arctic Sea the

[56] Cf. Sverdrup 1940.

freezing of surface water to form perennial sea ice would have been a factor[57] in reducing temperatures in the northern hemisphere. As the expanding Scandinavian Ice Sheet approached Britain the stormy conditions induced by its presence must have increased the precipitation of snow on the various British highlands, hastening the expansion of local glaciers in that region.

These and similar effects of the initial worldwide reduction of temperature induced, in specific regions, secondary reduction of temperature or increase of precipitation, or both. The processes involved were therefore self-extending in that they aided the expansion of glaciers by reinforcing their basic causes. At the present time it is not possible to determine what proportion of a temperature reduction as inferred from the geologic evidence is attributable to worldwide causes and what proportion of it was induced by secondary effects such as those mentioned.

Conclusion

We have now summarized the solar-topographic hypothesis. Its topographic aspect is well established on firm geologic evidence. Its solar aspect is consistent with the stratigraphic evidence, but the present-day observational evidence in support of it is limited to the fact of short-term fluctuation of the Sun's radiant energy through a small range. Beyond this small range the fluctuation has to be assumed. Nevertheless the hypothesis explains the seven items of basic data set forth at the beginning of this chapter. No other synthesis hitherto advanced in explanation of the Pleistocene climates does so. The idea has inherent probability, which will increase with any future increase in the observed range of fluctuation of solar energy. When the dates of the glaciations of the two polar hemispheres relative to each other have been established beyond argument, an important question concerning the firm basis of the hypothesis will have been answered. If the glaciations are thereby established as essentially contemporaneous throughout the world, the geologic basis of the solar-topographic hypothesis will be very firm indeed.

[57] C. E. P. Brooks (1936, p. 286).

Chapter 23

THE FOSSIL RECORD[1]

OUTSTANDING FACTS

Evidence of the life of the glacial and interglacial ages consists of fossil plants in peat bogs and buried soil layers, and fossil animal remains in caves, stream deposits, frozen ground, and oil seeps. Although fragmentary, this evidence, discovered little by little, is widespread. As a result the literature on Pleistocene life is very bulky, and many volumes have been written as reviews and summaries of knowledge in this field. The discussion that follows makes no attempt to deal with the stratigraphic range of animals and plants, a matter that is not only highly specialized but also, for the Pleistocene at least, difficult and uncertain. This discussion, rather, lays emphasis on some of the climatic and geographic implications of the Pleistocene fossils, a matter that is less uncertain.

So fragmentary is the evidence that even in the wide fields of climate and geography we can not go far. Yet the evidence appears to justify three broad generalizations. The first is that the Pleistocene floras and faunas of former times included most of the existing plants and mammals. Thus they were not radically different from those of today. However, they did include, in addition, a large number of mammal species that became extinct at various times after the maximum of the last glacial age, some of them in fact within the last few thousand years.

The second generalization is that in the cool-temperate zone in Europe, North America, and eastern Asia we find as fossils northern, cold-climate animals and, in addition, animals characteristic of warmer climates than now prevail in the regions where they are found. The northern group are believed to date from the glacial ages and the others from the interglacials. Thus there appear to have been widespread migrations in both directions, in response to slow climatic changes.

The third generalization is that, because of the occurrence of identical species of animals on land areas now separated by water, at one or more times during the Pleistocene epoch, North America had land connections

[1] For general discussions of Pleistocene fauna and flora see Baker 1920; 1929; Hay 1923; A. S. Romer 1933; 1945; Scott 1937. For Europe see Hopwood 1940. For early man see Nelson 1938; MacCurdy 1937; Howells 1944.

with Asia and with South America, England was connected with continental Europe, and various islands such as Sumatra and Borneo were united.

These generalizations are considered in more detail below.

COMPOSITION OF THE PLEISTOCENE FLORAS AND FAUNAS[2]

Virtually all the fossil plants of former Pleistocene ages are represented by living species. Apparently plants have undergone little change throughout the epoch. However, they did migrate as the climatic belts shifted equatorward and poleward with the waxing and waning of the glacial ages. The Pleistocene plant record therefore is a record of change of geographic position rather than of evolution of form.

Nearly half the reported species of Pleistocene fossil animals are mollusks, and most of these are land snails. A very large proportion of the fossil insects are beetles, probably because beetles are more readily preserved than other kinds of insects. The mollusks have undergone little change throughout the Pleistocene; in general the same kinds appear throughout the stratigraphic column. On the other hand the insects evolved rapidly. Apparently they were more sensitive as a group to the marked changes of climate that left the mollusks almost untouched.

The mammals, like the plants, have undergone little change throughout Pleistocene time, which appears to have been too short for conspicuous evolution to have occurred.

Contrary to earlier beliefs it has recently become evident that many if not most of the Pleistocene mammals in Europe appeared early in the epoch.[3] Groups of mammals, however, became extinct at various times during the Pleistocene. Thus a number of "old-fashioned" species which had flourished during the First Interglacial became extinct before the Second Interglacial.[4] These included *Hyaena arvernensis, Canis nescherensis, Equus robustus, E. stenonis, Rhinoceros etruscus,* and *Mammuthus (Archidiskodon) meridionalis.* During the Third Glacial these species became extinct: *Loxodonta (Paleoloxodon) antiquus, Rhinoceros megarhinus,* and *Felis leo* var. *spelaea.* Finally, late in the Fourth Glacial—in some cases very recently indeed—*Mammuthus primigenius, Coelodonta, Hyaena spelaea,* and *Ursus spelaeus* became extinct.

In North America the record is less abundant and less well dated

[2] The names of mammal genera and species used in this and other chapters follow in general the classification of G. G. Simpson (1945).

[3] Colbert 1942, p. 1475.

[4] Hopwood 1940, p. 86.

stratigraphically than the European record, especially as regards the earlier Pleistocene stages. However, it is becoming increasingly clear that in North America, contrary to earlier opinions, the mammals characteristic of the Pleistocene appeared at or near the beginning of the epoch and persisted throughout the Pleistocene.[5] A considerable group have become extinct virtually within the last few thousand years, whereas others still persist. The large mammals that died out include all the camels, all the horses, all the ground sloths, two genera of musk-oxen, peccaries, certain "antilopes," a giant bison with a horn spread of 6 feet, a giant beaverlike animal *(Castoroides)*, a stag-moose *(Cervalces)*, and several kinds of cats, some of which were of lion size. The "Columbian mammoth" *(Mammuthus [Parelephas] columbi)* and the "Imperial mammoth" *(Mammuthus [Archidiskodon] imperator)*, both larger than living elephants and common throughout the United States, also disappeared. The mastodon *(Mammut americanus)*, a forest dweller that ranged from Alaska to Mexico and eastward to the Atlantic coast, and the woolly mammoth *(Mammuthus primigenius)*, found throughout a broad belt lying just outside the margins of the ice sheets, chiefly in tundra and steppe environments, likewise persisted until very late in the Wisconsin age.

It is interesting to note that the teeth and bones of the mastodon are found most commonly in bog peat, whereas the remains of the woolly mammoth are not. This difference appears to reflect the difference between the habitats of the two animals as indicated above.[6]

It seems probable that many of these extinctions in both Europe and North America were the result of the hunting activities of prehistoric man. A number of the animals that died out are known to have been extensively hunted, and there appears to be no evidence of any climatic or topographic cause of extinction that meets the facts. To be sure, late-Pleistocene mountain uplift occurred in the Himalayan region and in the Alps, and large-scale rifting took place in eastern Africa. But in North America, where extinctions were widespread, significant crustal movements at this time are not apparent, except in coastal California. The idea that extinction resulted from human activity fits in with the fact that in Europe, where man lived throughout the Pleistocene, extinctions were progressive throughout the epoch, whereas in North America, apparently not entered by man until the Wisconsin age, extinctions did not begin until after man had arrived.[7]

[5] Colbert 1942, p. 1512; see also A. S. Romer 1933.

[6] Cf. Eiseley 1945. A. S. Romer (1933, p. 52) notes that reported occurrences of woolly mammoth in southern United States probably are based on incorrect identifications.

[7] Colbert 1942, p. 1513.

The problem of the extinction of large mammals by man was interestingly discussed by Pfizenmayer[8] and by Sauer,[9] who took the affirmative view, and by Eiseley,[10] who took the negative, preferring a hypothesis of extinction through climatic change. As yet the matter can. hardly be regarded as settled.

CLIMATIC IMPLICATIONS

The record of Pleistocene fossils as it applies to climatic inferences is peculiarly unsatisfactory. The record is in origin largely terrestrial rather than marine, in quantity small, and in character fragmentary. Many of the finds, in as well as outside the glaciated regions, are of unknown stratigraphic position. Furthermore the ecologic interpretations placed on most of the plants and animals in the record manifestly can never be more than approximate at best. In consequence few major inferences as to Pleistocene climates rest on fossil evidence alone. The chief value of fossil data is in strengthening inferences drawn from evidence of quite other kinds.

Another disadvantage in the use of fossil data is that, although the occurrence of boreal animals and plants in and near the areas covered by the great Laurentide and Scandinavian ice sheets indicates a cold climate, it is impossible to determine whether the climate resulted directly from worldwide reduction of temperature or whether it was in part a secondary effect resulting from the incursion of the ice sheets. It is very likely that the occurrence of abundant fossil spruce and fir, as well as of the subarctic woolly mammoth *(Mammuthus primigenius)*,[11] in central United States records merely the subarctic forest and the forest-tundra transition zone that spread out beyond the southern margin of the Laurentide Ice Sheet and that migrated with the ice sheet.

Conversely the occurrence in Pleistocene alluvium in central Alaska of a great cat *(Panthera atrox)* and a ground sloth *(Megalonyx)* indicate a climate warmer than now, but, as the fossils can not be assigned to a specific interglacial age, little more can be said about them at present.

In Alaska thus far the woolly mammoth has been found only in the

[8] Pfizenmayer 1939.

[9] Sauer 1944. For a more conservative view see A. S. Romer 1933, p. 76.

[10] Eiseley 1943.

[11] Bell (1898, p. 380) points out that, as the great length and curvature of the tusks of the mammoth were wholly unsuited to life in the dense subarctic forest, this animal probably kept largely to the forest-tundra transition zone. Probably this zone lay fairly close to the margin of the ice sheet throughout the greater part of each glacial age.

For a good general discussion of the woolly mammoth see Osborn 1930.

nonglaciated areas, not in the areas that have been glaciated. It has been inferred[12] that this animal was unable to survive the glacial ages in Alaska. The inference does not seem well founded, as the evidence is abundant elsewhere that the woolly mammoth lived close to the margins of large glaciers (though not actually *upon* glaciers as has sometimes been believed). This elephant was clearly an inhabitant of the forest/tundra transition zone, though also he ranged at times into much warmer climates. It is likely that, as the glaciated areas of Alaska are chiefly mountainous, erosion has removed from these areas much of the sediment in which mammalian fossils could be inclosed, and that this accounts in part at least for the distribution of the fossil finds.

The flora and fauna of the Toronto interglacial deposits (Chapter 14) are outstanding in that they record a mean temperature higher than now by 2° to 3° C. Similarly the fossils in the Hötting interglacial deposits in Austria, the Eem deposits in Denmark and northern Germany, and the deposits at Gerna and Bolnäs in middle Sweden and in various parts of European Russia[13] are alike in indicating climates milder than the climates now obtaining at those localities. Taken in conjunction with the evidence given by the lowered regional snowlines of mountain regions they indicate a fluctuation of mean temperature, between full glacial and full interglacial conditions, amounting in places to as much as 10° C.

On both eastern and western coasts of northern Italy there are peat deposits just above and just below sealevel, containing fossil spruce, fir, pine, birch, alder, willow, and possibly larch. In Italy today fir occurs only at high altitudes, and spruce and pine are restricted to the region of the Alps, 150 miles farther north. As these peat beds appear to have accumulated when sealevel was lower than now, it is thought that they date from one of the glacial ages and that they record a southward migration of Alpine forest trees with the colder climate of the glacial age.

The occurrence in western and central Europe of fossil mammals belonging to quite different climatic zones supports the evidence from other sources that the climatic belts shifted widely with the coming and going of the glacial ages.

In the glacial deposits of central and western Europe we find fossil lemming, arctic fox, reindeer, musk-ox, wolverine, and moose (all now living today in colder regions) and in addition the now-extinct but no

[12] Taber 1943, p. 1535. A list and description of the Alaskan mammals found and identified up to 1930 are given in Frick 1930.

[13] All these horizons are detailed in Chapter 16.

less cold-climate woolly mammoth and woolly rhinoceros which ranged southward into southern France. Among the plants we find fossil fir and spruce far south of their present ranges.

In the Crimea a fossil fauna of northern aspect has been found. It includes Arctic fox, northern deer, polar lark, white partridge, grouse, and chough, recording a far-southern penetration of this fauna during a glacial age.[14] Only one such fauna has been found as far south as the Crimea. It is believed by Russian authorities to date from the time of maximum glaciation of European Russia, when the Scandinavian Ice Sheet spread to within 200 miles of the Crimean peninsula.

On the other hand many mammals, probably interglacial, such as the jerboa, suslik, and saiga antelope, today living on the dry steppes of eastern Europe and Asia, are found as fossils in central and western Europe. Similarly such living species as the lion, hyaena, and hippopotamus as well as extinct species of warm-climate elephants and rhinoceroses occur as fossils in western Europe including Britain. Fossil rhododendron has been found high in the Alps. All these forms are believed to have occupied these regions, now too cold for them, during one or more of the interglacial ages.

In the steppe region of the Kazakh S.S.R. are Pleistocene deposits containing fossil plants and animals now foreign to this region and living in a moister climate, notably the oak tree and the peat-deer.[15]

Lake deposits in the Kharga Depression in Egypt contain leaves of an oak, *Quercus ilex,* that today does not grow nearer the equator than Corsica and southern France. Its presence in Egypt indicates a former temperature several degrees lower than that of today.[16]

In southern Peninsular India, certain isolated ranges of hills today support a flora and fauna including peculiar species that are found also in the Himalayas but are unknown throughout the broad intervening Indian plains. It is held that these species could only have migrated to southern India and have become isolated there as a result of a general reduction of temperature, which cooled the Indian plains sufficiently to encourage migration.[17]

In northern China the Chou-Kou-Tien (middle or upper Pleistocene) deposits have yielded an abundant and curiously mixed fauna. Tundra mammals including the woolly rhinoceros and lemminglike types, subarctic steppe mammals, and temperate-climate mammals including *Rhinoceros merckii* apparently lived contemporaneously in this

[14] Gromov 1945.
[15] Gerasimov and Markov 1939, p. 451.
[16] Hume and Craig 1911.
[17] W. T. Blanford, *quoted in* Wadia 1939, p. 277.

region.[18] This mixture might have arisen through the persistence of old faunal elements after a climatic shift accompanied by the arrival of invaders from a different ecologic zone. Such mixtures have not been found in Europe or in northern North America, perhaps because the presence of ice sheets brought about more pronounced climatic fluctuations than took place in northern China where no ice sheets formed.

The North American record is less clear than the European, but it, too, points to pronounced climatic swings. Reindeer and woolly mammoth ranged as far south as southern New England, moose inhabited New Jersey, walrus spread along the coast southward to Georgia, and musk-oxen ranged as far southwest as Kentucky, Arkansas, and Texas. Among the plants fir and spruce were common in Illinois and Iowa. All these occurrences are known or believed to date from the glacial ages.

Conversely fossil manatee has been found as far north as New Jersey, tapir and peccary occur in Pennsylvania, and panther and ground sloth occur in central Alaska. Near Toronto, Ontario, are fossil plants whose northern limits today are Ohio and Pennsylvania. All these occurrences appear to be interglacial.

In southwestern Kansas terrestrial deposits dating from the earlier part of the Pleistocene contain two successive assemblages of fossil rodents.[19] The older of these groups contains rodents living today in northern United States, western Canada, and Alaska. The existing relatives of the younger group are found in southwestern United States and Mexico. This evidence strongly suggests climatic fluctuation in Kansas from much colder than now to slightly warmer than now.

Colder climate is implied also by the fossil marmot found[20] in mountains near Santa Fe, New Mexico, at altitude 5900 feet. This species now lives slightly below timberline (11,000 ft.); hence a rise of timberline of at least 4000 feet since the time this marmot was alive is inferred. A rise of 4000 feet would correspond to an increase in mean temperature amounting to 6° C.

Santa Cruz Island (lat. 34°) off the coast of California is the site of a deposit containing a Pleistocene flora. This flora consists of nine species, which today reach their best development on the California coast 450 miles north of Santa Cruz Island. Comparison of the two localities indicates that, when these plants were growing on Santa Cruz, the island had a mean temperature 4° to 5° C lower than now, and a mean

[18] Lee 1939, p. 371.
[19] Hibbard 1944.
[20] C. E. Stearns 1942.

precipitation somewhat greater than now. The inference[21] is that this flora dates from the maximum of one of the glacial ages. Unfortunately there is no evidence as to which age is represented.

An assemblage of plants that occurs as fossils at Carpinteria, California, is living today on the California coast 200 miles farther north, where the climate is both cooler and moister than at Carpinteria.[22]

The upper part of the marine Santa Barbara formation is believed to be early Pleistocene. Exposures at San Pedro, California, include a considerable section containing fossil invertebrates. This assemblage today lives in water with a mean temperature of about 17° C. However, there is one zone containing invertebrates whose representatives today live in water at a mean temperature of only 10° C. Probably this "cold zone" represents one of the glacial stages.[23]

The invertebrate fauna of the Cape May-Gardiners clay formation along the Atlantic coast of the United States, supposed to belong to the Sangamon Interglacial stage, and described in Chapter 14, implies a climate slightly warmer than now.

In Tokyo Bay, Japan, calcareous Pleistocene sediments contain coral species that do not now live north of the Bonin Islands, a region with a mean annual temperature about 10° C higher than that at Tokyo. Hence it is inferred that the fossil corals at Tokyo flourished during an interglacial age when the temperature there was materially higher than it is today.[24]

In Australia various inferences have been drawn from fossil vertebrates. Pleistocene deposits at Darling Downs, Queensland, include an assemblage of fossil marsupials that could have flourished only when the vegetation of this district was distinctly more abundant than now. This would require greater rainfall and less evaporation. Again the occurrence of fossil crocodiles in alluvium near Port Augusta on the coast of South Australia, and of fossil lungfishes (*Epiceratodus*) east of Lake Eyre in central Australia, shows that when or just before these animals flourished at these places stream communication with tropical Queensland must have existed. The presence of continuous drainage in turn points to a climate distinctly more moist than the present one.[25]

The fossil occurrence in the Saharan region of North Africa of such

[21] Drawn by Chaney and Mason (1930, p. 24).

[22] Chaney and Mason 1930, p. 79.

[23] Crickmay 1929, p. 634. It has been shown (cf. Natland 1933) that distinctly different marine faunas can live in close proximity to each other, the habitat of each being controlled largely by water depths and temperatures. This fact must be considered before the occurrence of any distinctive fauna is ascribed to a change of climate.

[24] Olbricht 1923, p. 728.

[25] David 1932, p. 95.

mammals as *Mammuthus (Archidiskodon) meridionalis,* rhinoceros, hippopotamus, and crocodile surely record a stage of much greater rainfall than now occurs in that region.[26]

LAND BRIDGES

The fossil record indicates clearly that, at some time or times during the Pleistocene, North America had a land connection with Asia. Almost certainly this connection was in the vicinity of Bering Strait, which today separates Alaska from Siberia. Hardly more than 50 miles wide and very shallow, this Strait could have become dry land as a result of a 150-foot upwarping of this part of the Earth's crust or an equally moderate sinking of sealevel. The result would have been not a mere narrow bridge but a broad plain continuing as a fringe along the coasts of both Alaska and Siberia. Warmed by the Japan Current this plain would have had at least as mild a climate as that of the Aleutian Chain today. It is likely to have been covered with long thick grass like that now growing on the Alaska Peninsula—ideal fodder for the woolly mammoth and other herbivores. Even the dead grass beneath the winter blanket of snow should have provided nourishing forage.

Only by immigration over such a bridge does the presence in America of the woolly mammoth, musk-ox, bison, goat, moose, wapiti, and saiga antelope (to name only a few common animals) seem explicable. At some time, probably at various times, during the Pleistocene these and other animals entered North America from Asia.

Among the Pleistocene fossil faunas in North America we find also South American animals such as ground sloths occurring as far north as central Alaska. Conversely we find in South America horse and deer, which can only have reached that continent from North America. It is highly probable, therefore, that at times if not continuously during the Pleistocene the present Isthmus of Panama was in existence. There can be little doubt that this land bridge was the route of intermigration. Today this land bridge is marked by a 200-mile belt of swampy tropical rain forest. This would be no barrier to an animal such as the peccary, which did in fact migrate from South to North America, apparently early in the Pleistocene epoch. But it would have been a serious obstacle to the animals named above. It would have been an obstacle also to the intermigration of grasses, shrubs, and herbs characteristic of semiarid environments, which also succeeded in spreading from one continent to the other at this time. It has been suggested that the explanation is to be

[26] C. E. P. Brooks 1932, p. 88.

found in the southward shift of climatic belts that took place during the glacial ages and that brought on pluvial climates in now-arid southwestern North America. A shift of the northern extratropical belt of high pressure southward through only 2 degrees of latitude should have dried the isthmus sufficiently so that the rain forest would have been replaced by savanna. This would have opened the isthmus to travel by plants and animals that could not easily, if at all, have crossed the tropical forests.[27]

This hypothesis implies that the migrations took place only during glacial ages, for only at such times could the climatic belts have been pushed toward the equator. The lowered sealevels of the glacial ages should have aided migration rather than hindered it.

Shifts of sealevel, however, are believed to have been a cause of the linking of Britain with the continent of Europe. The English Channel, now barely 100 feet deep at its narrowest part, must have been replaced by an isthmus during the lowered sealevels of the glacial ages. That there was a similar land connection during certain interglacial times is very strongly suggested by the presence in Britain of fossil mammals of warm-climate types such as *Mammuthus (Archidiskodon) meridionalis, Hippopotamus,* lion, and hyena.

Early in the Pleistocene, at least, a land connection existed between Japan and the Asiatic mainland, because several species of mainland mammals, especially proboscideans, occur in the lower Pleistocene deposits of Japan.[28] A lowering of present sealevel of less than 200 feet would create a land bridge from the lower Amur River region in Siberia via Sakhalin Island to Japan. However, as the land connection existed also in pre-Pleistocene times, the emergence need not have resulted from glacial lowering of sealevel.

The Sunda Sea between Borneo and Sumatra is shallow enough to have been converted into a land bridge during the glacial ages. In this way are explained the many similarities between the living terrestrial and fresh-water faunas of Borneo and Sumatra.[29]

Similarities among the fossil mammalian faunas of Australia, Tasmania, and New Guinea suggest that at times during the Pleistocene free migration between these lands was possible. Bass Strait, now separating Tasmania from Australia, could be made a land bridge by a lowering of the sealevel by 180 feet, and a slightly greater lowering would make a land bridge of the strait that now separates Australia from New Guinea.[30]

[27] Sauer 1944, p. 557.
[28] Colbert 1942, p. 1490.
[29] Cf. Molengraaff and Weber 1921.
[30] Data on the Pleistocene fauna and flora of Australia are summarized in Browne 1945.

EARLY MAN IN THE OLD WORLD

As far as we now know, the geologic history of man, as distinct from the apes, is confined largely or entirely to the Pleistocene epoch. The human (and prehuman) record is still very scanty, but, such as it is, it has been pieced together chiefly from two kinds of evidence. The first consists of skeletons or parts of skeletons, which it is one of the objectives of anthropology to discover and study. The other kind of evidence consists of *artifacts*—implements of stone, bone, and the like—and drawings and carvings on the walls of caves and on pieces of ivory and other materials. It is part of the task of archeology to unearth and study these objects.

The skeletal parts show that marked evolution took place during the million-year stretch of Pleistocene time, particularly in the brain, which increased greatly in size. The artifacts show a general progressive increase in perfection and adaptability, which in turn record an increase in intelligence among the people who made them. These increases, however, apparently took place at different rates in different regions; very different degrees of culture flourished at the same time in different parts of the world, as is still true today. This has been learned in part through dating the various groups of artifacts and skeletal remains, both by means of fossil mammals that have been preserved with them, and by identifying the deposits, in which some of the remains occur, with deposits of known glacial or interglacial age. The much-generalized result of this work is shown in Table 30.

In this table six principal types of fossil man preceding modern man are shown, as are four principal culture stages (based on perfection of artifacts) prior to the Age of Metals. Several distinct and widely recognized subdivisions of the Paleolithic stage (also known as the Old Stone Age) are shown also. The culture stages have been worked out from finds in Eurasia and Africa. The record in America does not extend as far back in time as that in the Old World. Also it consists largely of artifacts, for skeletal remains dating from times earlier than very late Wisconsin have not yet been found in America.

The table is necessarily somewhat artificial, because cultural evolution did not keep pace everywhere with organic evolution. Such terms as "Paleolithic" and "Neolithic" are unavoidably confusing because they are not strictly time terms; they refer to types of culture which overlapped in time. In general the crude objects referred to the Eolithic stage are of doubtful human origin; opinion is divided as to whether they were made by early man or by natural agencies. Paleolithic culture is marked by flaking and chipping of flint. Neolithic culture is marked, in addition, by pecking, grinding, and polishing of various kinds of stone.

TABLE 30. GENERALIZED CORRELATION OF THE HUMAN RECORD IN EURASIA WITH THE GLACIAL RECORD
(Modified after N. C. Nelson and others)

Geologic Time Units			*Dates in Years* B.C. *(estimated)*	*Cultural Stages*			*Types of Fossil Man*	
	North America	*Europe*						
Pleistocene epoch	Wisconsin Glacial age	Fourth Glacial age	5000 6500	Age of Metals Neolithic Mesolithic			*Homo sapiens*	Modern man
Pleistocene epoch	Wisconsin Glacial age	Fourth Glacial age	15,000 55,000	Paleolithic	Late	Azilian Magdalenian Solutrean Aurignacian	*Homo sapiens*	*Cro-Magnon man*
Pleistocene epoch	*Sangamon Interglacial age*	*Third Interglacial age*	100,000 225,000	Paleolithic	Middle	Mousterian		*Neanderthal man*
Pleistocene epoch	Illinoian Glacial age	Third Glacial age	325,000	Paleolithic	Early	Acheulian Levalloisian		
Pleistocene epoch	*Yarmouth Interglacial age*	*Second Interglacial age*	600,000	Paleolithic	Early	Chellean Choukou-tienian Clactonian		*Heidelberg man* *Peking man* *Pithecanthropus* *Swanscombe man*
Pleistocene epoch	Kansan Glacial age	Second Glacial age	700,000	Paleolithic	Early			
Pleistocene epoch	*Aftonian Interglacial age*	*First Interglacial age*	900,000	Paleolithic	Early	Pre-Chellean		*Piltdown man*
Pleistocene epoch	Nebraskan Glacial age	First Glacial age	1,000,000+	Paleolithic	Early			
Pliocene epoch				Eolithic				

EARLY MAN IN THE NEW WORLD[31]

It was long believed that man was a very recent arrival in the New World — that in the Americas the first comers were the Indians who barely antedated modern times. This belief grew up at a time when the succession of glacial and interglacial ages was very imperfectly understood, when the fluctuations of the Wisconsin glaciers were almost wholly unknown, and when the absolute chronology of Pleistocene events had only been guessed at. In consequence archeologists felt themselves forced to place the arrival of man (apparently the modern Indian) in America at a time after the latest great ice sheets in middle latitudes had largely or entirely disappeared. The only apparent alternative, that man had turned up in North America during the Sangamon Interglacial age, would put the cultural succession in America so far ahead of that in Europe that it was not to be thought of.

The growth of knowledge concerning climatic fluctuations during the Wisconsin age, coupled with new finds of artifacts that are demonstrably prehistoric even in Old World terms, has changed radically the former point of view. It is now established that man has inhabited America for at least 20,000 years and possibly for 40,000 to 60,000 years.

The most important factor in the modern view is the find made in 1927 near Folsom, in northeastern New Mexico, of a human culture associated with the bones of extinct mammals. The artifacts and bones occur in lake deposits where no lake now exists. Thus the climate appears to have been pluvial, and pluvial conditions in that region must have antedated considerably the Climatic Optimum. After the principal facts concerning this culture,[32] widely known as the *Folsom culture,* were established, still earlier cultures were discovered in various parts of western United States.

A considerable array of differing cultures in the Americas is now recognized. At least two pre-Folsom cultures have been identified from sites in Texas, New Mexico, and California. These were followed by the Folsom culture, characterized by skilfully grooved bayonet-shaped spear points made of flint. These "Folsom points" have been found at various localities from the Mexican border through the Great Plains of the United States and Canada to Alaska.

The Yuman culture, marked by fine, nongrooved, leaflike blades, apparently was contemporary with the Folsom but outlasted it by a long stretch. All these cultures — pre-Folsom, Folsom, and Yuman — are those of people who lived mainly by hunting. Their quarry consisted of the

[31] See a good summary in Nelson 1933.

[32] MacCurdy 1937, pp. 139–152.

woolly mammoth, the mastodon, the extinct bison, and other extinct mammals. The Yuman people were followed after a long interval by later groups (known as the Basket Makers, the Hopewell People, and others) who were farmers as well as hunters. These in turn were followed by the Indians of modern times. In saying "followed" we do not imply that the Indians of historic time are comparatively late immigrants. Rather they seem to be for the most part the results of various combinations and mixtures of the earlier peoples outlined above.

Except for the more recent groups the early American peoples are known only from artifacts and similar evidence and are represented doubtfully or not at all by skeletal remains. The lack of skeletal evidence is one of the factors that held back until recently a widespread belief in the great antiquity of man in America. The explanation of the scarcity of skeletons is not known. It may lie partly in less thorough search for them — there has been less time for search — than in Europe. It may lie partly in much less dense prehistoric populations in America than in Europe. To some degree it may lie in customs of disposing of the dead either by abandonment or by deliberate exposure above ground, as is practiced by some Indian tribes today, rather than by burial.

The absolute dating of the various ancient American peoples is still largely a matter of conjecture. Folsom man dates back at least 20,000 years. Pre-Folsom man may have an antiquity of 40,000 to 60,000 years; that is to say, he may possibly date from as far back as the Iowan sub-age.

Paths of Immigration

Finds of Folsom points in western and central Alaska suggest strongly that the Bering Strait was the route by which the Folsom people reached North America from Asia. It seems likely that the pre-Folsom peoples had made use of the same path. Whether these pioneers crossed by a land bridge such as might have existed during the lowered sealevel of the Iowan sub-age, or whether, as seems more likely, they made the journey over ice such as forms a continuous cover over the strait in exceptionally cold winters nowadays, probably we shall never know. Central and western Alaska were not covered by glaciers, so that in those regions living conditions should have been about as good as they have been for the native peoples in the recent past. It is likely that during the warmer sub-ages both forage and game were plentiful.

The path taken by the succession of primitive peoples from Alaska to the heart of North America is purely conjectural. The map (Plate 3) showing the extent of former glaciers suggests that, at least during the times of partial deglaciation in the Wisconsin age, relatively low-level

easy routes should have been open southeastward through the interior of British Columbia and, farther east, along the east base of the Rockies, roughly the route of the modern Alaska Highway.

There seems to be little doubt that South America was peopled from North America via the Isthmus of Panama and in part by direct immigration from Pacific islands.

SUMMARY

In summary, the mammalian fauna of the present day differs only in detail from that which has prevailed throughout the greater part of the Pleistocene. However, the whole epoch has been marked by extensive migrations of both plants and animals, including man. Well-trod paths of migration have included the isthmus connecting the two Americas and the narrow strait that today separates North America from Asia. The prime cause of the repeated migrations can not have been other than the shifting of the climatic belts as the climates slowly fluctuated from glacial to interglacial and back again. These migrations were but one of several responses to the broad irregular fluctuations of climate. As the climate grew cold and then warm the ice sheets formed, expanded, and then shrank away, areas of the Earth's crust beneath them were pushed down and then recovered, and the level of the sea was drawn down and then rose. Thus man played his small part in a vast synchronous series of actions and reactions in which the atmosphere, the sea, and the Earth's solid crust responded each in its own majestic way.

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